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Paul J. Kocin
Louis W. Uccellini

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Laboratory for Atmospheres
National Aeronautics and
Space Administration
Goddard Space Flight Center
Greenbelt, MD 20771

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ABSTRACT

Surface and upper-level characteristics of selected meteorological fields are summarized for eighteen major East Coast snowstorms dating from 1960 to the present. Two major "types" (Miller, 1946) of sea-level development are described and applied to the cases at hand, with a few storm systems showing characteristics of both types. Type "A" storms typically cross the Gulf Coast states and propagate northeastward along the Atlantic coast. They exhibit no secondary redevelopment of the surface low, although there may be a tendency for the low to "jump" northeastward along the coast. Type "B" storms typically approach the Appalachian Mountains from the west or southwest and then redevelop further to the south and east near the Atlantic coast. Aspects such as rapid sea-level deepening, coastal frontogenesis, cold air damming, low-level jet formation, the development of an "S"-shaped isotherm pattern at 850 mb, diffluence downwind of a negatively tilted upper-level trough axis, upper-level confluence across the northeastern United States, and an increase of geopotential heights at the base of the upper-level trough characterized the pre-cyclogenetic and cyclogenetic periods of many of the storm systems. However, large case-to-case variability was also observed, especially with regard to the spatial dimensions of the surface and upper-level systems, as well as variations in trough/ridge amplification and the evolution of upper-level jet streak systems. The influence of transverse circulations associated with a confluent jet streak entrance region in the northeastern United States and the diffluent exit region of a jet streak/trough system approaching the East Coast on the production of snowfall is also discussed.

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1. INTRODUCTION

Heavy snow accumulations and strong winds associated with extratropical cyclones along the East Coast of the United States during the winter season have had crippling effects on some of the nation's largest metropolitan areas. Snowstorms in urban environments can overwhelm snow-removal operations, restrict movement and severely curtail business, resulting in social and economic hardship for millions of people (see Cochrane et al., 1975). As the following two examples illustrate, the impact of such storms can be staggering. In the Boston, Ma. metropolitan area, as well as much of southern New England, most business and travel were suspended for a week or more following a storm that dropped 60 to 90 cm of snow that was drifted severely by surface winds gusting in excess of 35 m s^{-1} on 6-7 February 1978 (Brown and Olson, 1978). On 11-12 February 1983, the Washington, D.C., Baltimore, Md., Philadelphia, Pa., New York City, N.Y., and Boston, Ma., metropolitan areas were the sites of a major snowstorm that produced 40 to 60 cm accumulations (Sanders and Bosart, 1985a,b; Bosart and Sanders, 1985). This cyclone immobilized dozens of smaller urban centers as well, affecting more than 20 million people.

Extratropical cyclogenesis has long been the topic of extensive research that has taken several approaches, including the pioneering work of the Norwegian School that formulated the "Polar Front" theory of cyclones (Bjerknes, 1919; Bjerknes and Solberg, 1923), the application of "Dines compensation" to surface cyclogenesis (Bjerknes and Holmboe, 1944; Bjerknes, 1951a; Reiter, 1963), instability theories of cyclone growth

(e.g., Charney, 1947; Eliassen, 1956), and quasi-geostrophic approaches to cyclogenesis and associated vertical motions (Sutcliffe, 1947; Sutcliffe and Forsdyke, 1950; Petterssen, 1955). See Petterssen (1956), Palmen and Newton (1969), and Holton (1979) for textbook treatments of cyclogenesis.

The study of East Coast winter storms, in particular, has attracted considerable attention in the past, especially during the 1950's (Smith, 1950; McQueen and Keith, 1956; Spar, 1956; Mook and Norquest, 1956; Fuge and Kipper, Jr., 1957; Sanderson and Mason, Jr., 1958). In recent years, these storms are receiving renewed interest, serving as a major topic at annual workshops that examine many aspects of cyclogenesis. The storms are also a motivation behind the Genesis of Atlantic Lows Experiment (GALE) scheduled for 1986. The revived interest may be due to several recent, widespread, major snow events in the large urban centers that span the coastal northeastern United States, most notably on 19-20 January 1978, 6-7 February 1978, 18-19 February 1979, 6-7 April 1982 and 11-12 February 1983, after a period of years in which major snowstorms were virtually absent. It is important that these storms be studied and understood since public perceptions of forecast accuracy are particularly acute during these events due to their widespread impact on large segments of the population and commerce.

Operational and research numerical weather forecasts of East Coast cyclogenesis have performed well in many of these and other situations, such as the February 1978 cyclone (Brown and Olson, 1978) and the spring snowstorm of 6-7 April 1982 (Kaplan et al., 1982), but have failed on other occasions, most notably the 18-19 February 1979 "Presidents' Day" cyclone (Bosart, 1981; Newell, 1981; Uccellini et al., 1982; Uccellini et al.,

1983; Uccellini et al., 1984; Atlas, 1984). While the recent 11-12 February 1983 snowstorm and its associated precipitation amounts were, in general, predicted well by the National Meteorological Center's (NMC) Limited Fine Mesh (LFM) model, a small error in the model's forecast of the northward limit of heaviest precipitation amounts had likely resulted in the underprediction of significant snowfall in New York City and southern New England.[1] Therefore, even a relatively good numerical forecast may have small errors that can have a significant impact on the general population. The inconsistencies of the model forecasts of these cyclones suggest that certain dynamical features or processes associated with cyclones may be better simulated or diagnosed than others. Bosart (1981) and Uccellini et al. (1981, 1984) have noted some of the deficiencies of the models, which include inadequate boundary layer and cumulus parameterizations, deficient vertical resolution and poor jet streak simulations. However, recent studies using mesoscale models have shown that those models show promise, providing detailed forecasts of East Coast cyclogenesis, including the secondary redevelopment of surface lows east of the Appalachian Mountains (Kaplan et al., 1982; Zack et al., 1984; Kocin et al., 1985).

While the difficulties of predicting the movement and intensity of such storms are problems shared with many sections of the nation, topographical influences on the low-level cyclonic structure, and ultimately the entire cyclonic circulation, favor the East Coast as a primary location for significant cyclonic development (Reitan, 1974;

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[1]National Weather Service forecasts for New York City were obtained from the New York City forecast office at Rockefeller Center.

Colucci, 1976; Sanders and Gyakum, 1980; Roebber, 1984; and others). These topographical factors include land-sea temperature contrasts, the location and intensity of the Gulf Stream, air mass modification over the Atlantic Ocean, Great Lakes, and the Gulf of Mexico (which include the effects of latent and sensible heating), land-sea frictional effects, the shape of the coastline and the influences of the Appalachian Mountains. The ability of models to incorporate these effects is crucial to the successful forecasting of snowstorms along the East Coast.

The purpose of this study is to describe the characteristic surface and upper-level patterns of selected wind, pressure and temperature fields for a sample of some of the most significant East Coast cyclones since 1960. These patterns are examined from a survey of surface, 850, 700, 500, 300 and 200 mb charts derived from NMC analyses available on microfilm from the National Climatic Center (NCC) in Asheville, N.C. Eighteen cases, which serve as the basis for this study, were selected because they produced crippling snow accumulations across one or more of the urban centers spanning the Washington, D.C. to Boston, Ma. corridor (Table 1) and were associated with large areas of accumulations exceeding 25 cm from the Middle Atlantic states to New England (Fig. 1). Three of the cases occurred in December, four in January, eight in February, two in March and one in April. Two of the cases, December 1969 and February 1972, were selected for their widespread impact, although heaviest snow fell immediately north and west of the urban centers. A study of heavy snow systems that affect the East Coast was performed by Brandes and Spar (1971) but their composite analyses could not resolve antecedent patterns 12 to 24 h prior to the onset of heavy snowfall in various regions along the East

Table 1
Storm Dates and Snowfall Amounts

Storm Date		Washington, D.C.	Baltimore, Md.	Philadelphia, Pa.	New York, N.Y.	Boston, Ma.
		Snowfall in inches (cm)				
MAR 2-4	1960	8 (20)	10 (26)	8 (21)	16 (39)	20 (50)
DEC 10-13	1960	9 (22)	14 (36)	15 (37)	17 (44)	13 (33)
JAN 18-21	1961	8 (20)	8 (21)	13 (34)	10 (25)	12 (31)
FEB 2-5	1961	8 (21)	11 (27)	10 (26)	24 (61)	14 (37)
MAR 4-7	1962	4 (10)	13 (33)	7 (17)	1 (2)	1 (2)
JAN 11-14	1964	10 (26)	10 (25)	7 (18)	13 (32)	9 (23)
JAN 28-31	1966	14 (35)	12 (31)	8 (21)	7 (17)	6 (16)
DEC 23-25	1966	9 (23)	9 (22)	13 (32)	7 (18)	6 (14)
FEB 5-8	1967	12 (30)	11 (27)	10 (25)	15 (39)	10 (24)
FEB 8-10	1969	5 (13)	3 (8)	3 (7)	20 (51)	11 (28)
FEB 23-28	1969			2 (5)	2 (4)	26 (67)
DEC 25-27	1969	12 (31)	6 (15)	5 (13)	7 (19)	4 (11)
FEB 18-20	1972	10 (25)	3 (8)	4 (9)	6 (16)	6 (16)
JAN 18-21	1978	8 (19)	6 (14)	13 (34)	14 (36)	22 (54)
FEB 5-7	1978	2 (6)	9 (23)	14 (36)	18 (45)	27 (69)
FEB 17-20	1979	19 (47)	20 (51)	14 (36)	13 (32)	
APR 5-7	1982			4 (9)	10 (24)	13 (34)
FEB 10-12	1983	23 (58)	23 (58)	21 (54)	22 (56)	14 (34)

Table 1. Dates of East Coast snowstorms and snowfall amounts in inches (and cm) for Washington, D.C. (the maximum amount at either Dulles International Airport or National Airport), Baltimore, Md., Philadelphia, Pa., New York, N.Y. (the maximum amount at either Central Park, LaGuardia Airport, Kennedy International Airport, or Battery Place), and Boston, Ma.

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STORM SNOWFALL
IN EXCESS OF 25 CM

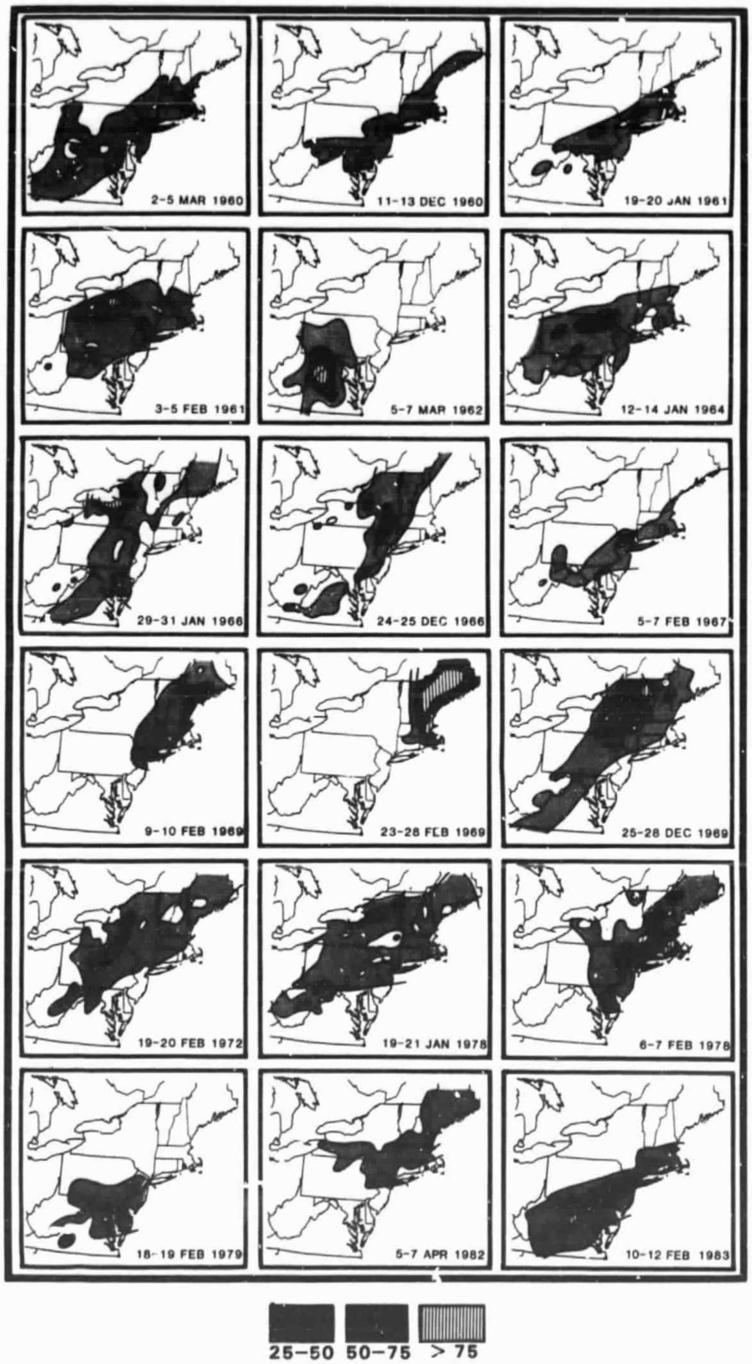


Fig. 1. Storm snowfall in excess of 25 cm for eighteen East Coast snowstorms since 1960 (light shading, 25 to 50 cm; heavy shading, 50 to 75 cm; linear shading, greater than 75 cm).

Coast. Their compositing approach may have smoothed out the detail and variability inherent in the pre-cyclogenetic environment that characterizes East Coast snow situations. Only a few composites are attempted in this study since the case-to-case variations of the most notable storms preclude effective compositing, especially for upper-level fields.

The quasi-geostrophic aspects of these cyclones, in terms of vorticity and thermal advections, are addressed only briefly in the text. A more thorough analysis for the eighteen cases requires a vast amount of time and funding for the data preparation needed to compute the quasi-geostrophic fields. The analysis of selected height, wind, and temperature fields presented in Parts I and II does, however, lend insight into the quasi-geostrophic behavior of these systems.

The study is divided into a summary of surface and upper-level characteristics (this paper) and a collection of case studies that contains detailed map discussions for each of the eighteen cases listed in Table 1 (Kocin and Uccellini, 1985; hereafter referred to as Part 2). Six-paneled 12-hourly surface, 500 mb (with 300 and 200 mb wind analyses) and 850 mb analyses, infrared satellite images (available only for the most recent cases), and a total snowfall chart were constructed for each case and cover a 60 h period which includes the pre-cyclogenesis, cyclogenesis and occlusion stages of most storms. Since the surface analyses were chosen to coincide with the 12 h availability of the rawinsonde data (either 0000 GMT or 1200 GMT), the ability to consistently portray a similar feature over the same geographical area for every case is limited. It is not the intent of this paper to provide objective forecasting guidance for heavy snow in the Eastern region, such as that described by Penn (1948), Bailey (1960),

Goree and Younkin (1966), Browne and Younkin (1970), and Spiegler and Fisher (1971). Rather, general descriptions of key dynamical patterns are provided for a large selection of cyclones to emphasize the case-to-case variability and similarities in snowstorms which influence some of the nation's most populated centers. Overviews of surface and upper-level characteristics are provided in Sections 2 and 3, respectively, and results of the study are summarized in Section 4. The case-by-case descriptions in Part II should be of use to both the forecaster and researcher who wish to diagnose or forecast the various processes which interact to produce major East Coast snowstorms.

2. OVERVIEW OF SURFACE CHARACTERISTICS

The surface weather characteristics of the eighteen cyclones are summarized in Table 2. The storms are first represented in terms of a classification proposed by Miller (1946) that describes two different "types" of sea-level low pressure development. The paths and speeds of the sea-level low pressure centers, the occurrences of coastal frontogenesis (Bosart et al., 1972) and cold air "damming" (Richwien, 1980), and deepening characteristics are also discussed.

a. Cyclone Type

The surface characteristics of East Coast storms have been described by Austin (1941), Miller (1946), and Mather et al. (1964), but Miller's study is the most extensive, having examined two hundred cases spread over a ten-year period ending in 1939. Miller categorized cyclones developing along the East Coast into two types using the surface weather map as the basis for the classification. Type "A" cyclones form along the frontal boundary separating an outbreak of cold, continental air located to the north-northwest of the incipient cyclone center from warmer, maritime air on the other side of the surface front. Type "B" cyclones reflect a more complicated sea-level development as a secondary low pressure center develops to the southeast of an occluding primary surface low pressure center, which is located typically over the Ohio Valley. The secondary low pressure center develops on the warm front of the primary system along a boundary separating a shallow, cold wedge of air to the east of the Appalachian Mountains and to the west of the coast from warmer air, which

Table 2
Summary of Surface Characteristics

Date	Cyclone Type (Miller, 1946)	Average Propagation Rates ($m\ s^{-1}$) primary	secondary	Stalling	Coastal Frontogenesis /Damming
M 1960	B	11-21	11-15	X	X
D 1960	B	9-16	16-17		X
J 1961	A	14-20			
F 1961	B	11-12	11-12	X	X
M 1962	AB	discontinuous	<12	X	
J 1964	B	9-10	9-12		X
J 1966	A	10-25		X	X
D 1966	A	10-15		X	
F 1967	A	17-28			
F 1969a	B	12	6-20	X	X
F 1969b	AB	3-19		X	X
D 1969	A	5-22		X	X
F 1972	B	6	7-16	X	X
J 1978	A	8-20			X
F 1978	AB	5-12		X	
F 1979	AB	14-15			X
A 1982	B	10-20	10-13		X
F 1983	A	10-19		X	X

Table 2. Summary of storm characteristics derived from 60 h surface analyses, including storm type according to Miller (1946); average propagation rates ($m\ s^{-1}$) of both primary and secondary low pressure centers; whether "stalling" occurred (forward motion of storm slowed by at least 5 $m\ s^{-1}$); and the presence of coastal frontogenesis/cold air damming.

is advected northward along the coastline. Miller also notes that several cyclones have development characteristics of both types.

Miller's classification is used in Table 2 to simplify a description of surface cyclonic features. Of the cases examined, seven are classified as type "A" systems, seven as type "B" and four cases have characteristics of both "A" and "B" (type "AB"). It should be noted that Miller's classification was defined for cyclones that formed near the Atlantic Coast (see Fig. 6, p. 36). Only nine of eighteen cases selected here, and none of the type "A" storms, formed within the domain that Miller chose for his study. As Miller points out, type "A" storms that form near the East Coast frequently pass out to sea without much effect on land areas. While Miller's study did not include East Coast cyclones that formed at locations distant from the Atlantic Coast, his approach is general enough to apply to all the cyclones studied here.

In this study, the seven type "A" cyclones responsible for heavy snowfall in the Northeast urban region typically developed or propagated across the Southern Plains states or the Gulf of Mexico (see Fig. 2a) and then moved northeastward along the Atlantic Coast, passing generally off the Carolina coastline. In contrast, the primary low centers of the seven type "B" cyclones (Fig. 2b) took widely varied paths, some moving northeastward from the Gulf of Mexico and others moving eastward or southeastward from a region extending from Texas to the Great Lakes. Most of the primary low centers terminated to the west of the Appalachian Mountains in the eastern Ohio Valley. The secondary low centers (Fig. 2b) formed generally from the northeastern Gulf of Mexico through Georgia, South Carolina, North Carolina, southeastern Virginia, and off nearby

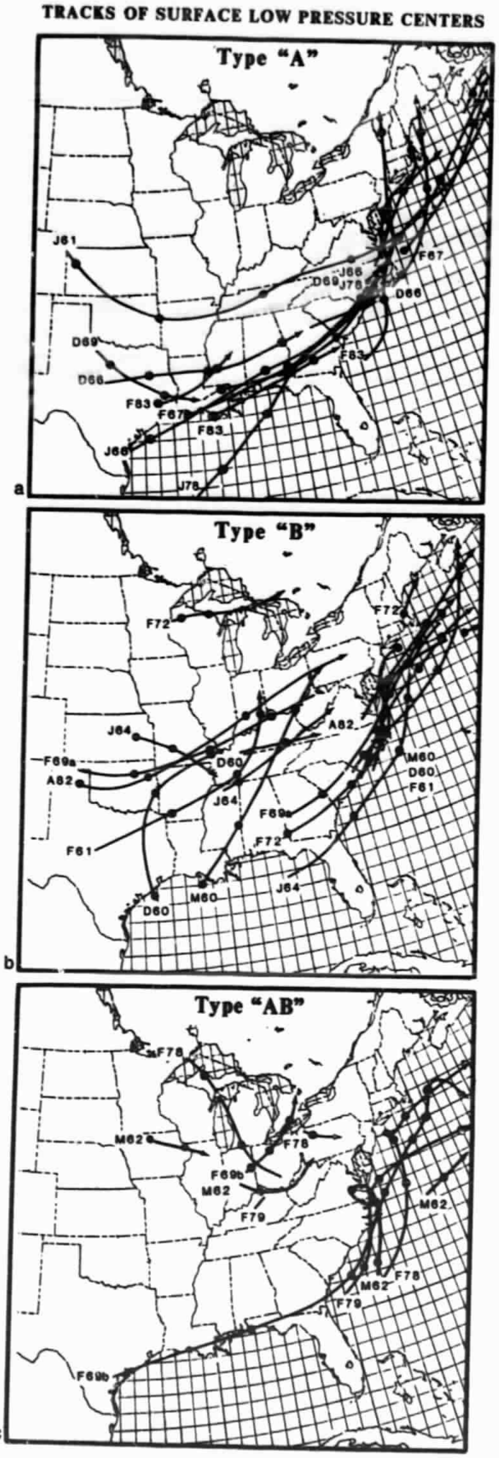


Fig. 2. Tracks of surface low pressure centers over the 60 h study period, grouped according to Miller's (1946) classification (type "A", type "B", type "AB"), including 12-hourly positions.

shores. Nearly all of the type "A" and the secondary type "B" cyclones consistently tracked northeastward from the Virginia coast to off southeastern New England, remaining approximately 100 to 300 km offshore. Similar results for heavy snowfall in New York City were found by Spar et al. (1967).

The four cases with the double classification (Fig. 2c) showed characteristics of both type "A" and type "B." Each case could be considered type "B" development since two separate low centers were present. However, the "primary" low was either weak (not occluding) or associated with poorly defined or no surface fronts, or the "secondary" surface low appeared to have developed independently of the "primary" low (i.e., there was no warm front associated with the "primary" low on which the "secondary" developed). In each of these cases, a surface low moved north to northeastward up the Atlantic Coast, also within 100 to 300 km of the coastline.

Examples of type "A" and type "B" storms are presented in Fig. 3. The rapidly moving February 1967 storm that produced brief, but paralyzing, blizzard conditions from Washington, D.C. to Boston, Ma., is shown in Fig. 3a to portray type "A" development. In this example, the surface low developed by 1200 GMT 6 February near the Texas coast along the leading edge of a major arctic outbreak. The surface low moved eastward across the Gulf Coast and then rapidly northeastward to a position off the Virginia coast by 1200 GMT 7 February. No secondary low developed in this case, but the cyclone did exhibit a tendency to "jump" or redevelop further northeastward along its path, as did several other type "A" cases. The January 1961 "Kennedy Inaugural" snowstorm is a relatively unique type "A"

TYPE "A" TYPE "B"

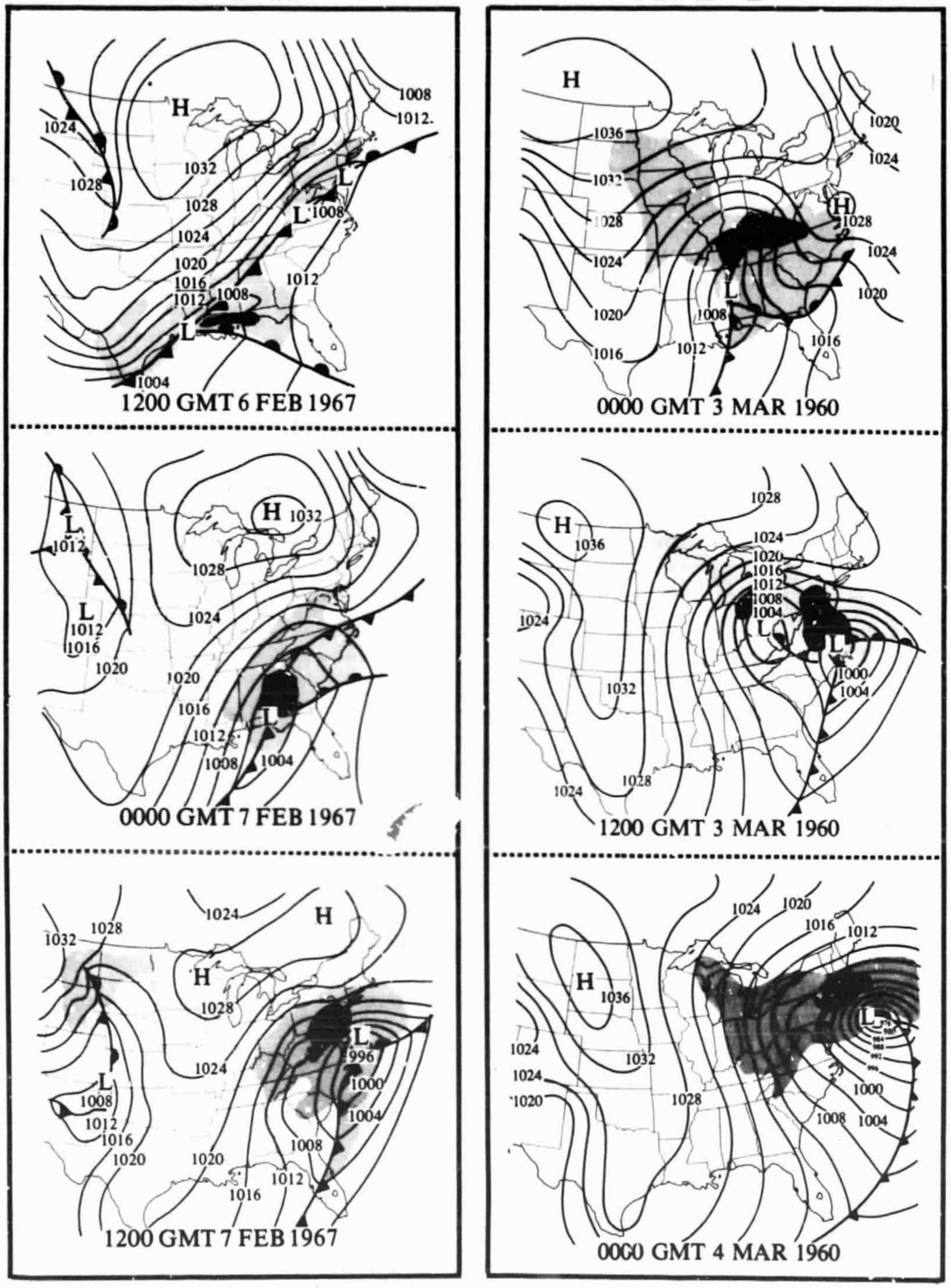


Fig. 3. Examples of type "A" and "B" (Miller, 1946) cyclones using 12 h surface analyses. Charts include frontal analyses, isobars (mb), high and low pressure positions and precipitation (shaded; heavy shading denotes moderate to heavy precipitation).

system since it followed a path more typical of type "B" cyclones (see Fig. 2), but no secondary low formed. Rather, the primary low center deepened as it moved rapidly across the Tennessee Valley to the Virginia coast. Its rapid rate of movement possibly inhibited or masked the redevelopment process.

The March 1960 snowstorm produced 50 to 75 cm accumulations and severe blizzard conditions across southeastern New England, and is shown in Fig. 3b to illustrate type "B" development. The primary low formed along the Gulf Coast prior to 0000 GMT 3 March and moved northeastward toward Ohio, where it deepened slowly. As the cyclone moved to the west of the Appalachians, a secondary low formed to its southeast across South Carolina and moved rapidly northeastward to off the Virginia coast by 1200 GMT 3 March. As the secondary low deepened rapidly off the Middle Atlantic coast, the primary low filled and was unidentifiable by 0000 GMT 4 March.

While only two "types" of surface characteristics have been ascribed to the storms presented in this study, their paths, size, intensity, precipitation distribution, frontal and sea-level pressure configurations varied considerably from case to case. Nevertheless, the surface development of a few storms resembled those of other cases, such as the February 1979 "Presidents' Day" cyclone and the February 1961 storm (Fig. 4), indicating that similar processes may have been operating in both systems. The 1979 "Presidents' Day" snowstorm, which paralyzed portions of northern Virginia, Washington, D.C., Maryland, Delaware, and New Jersey, may be categorized as either "A" or "B" since the primary low in the Ohio Valley was not associated with surface fronts prior to the rapid development of a separate secondary low off the East Coast on 19 February

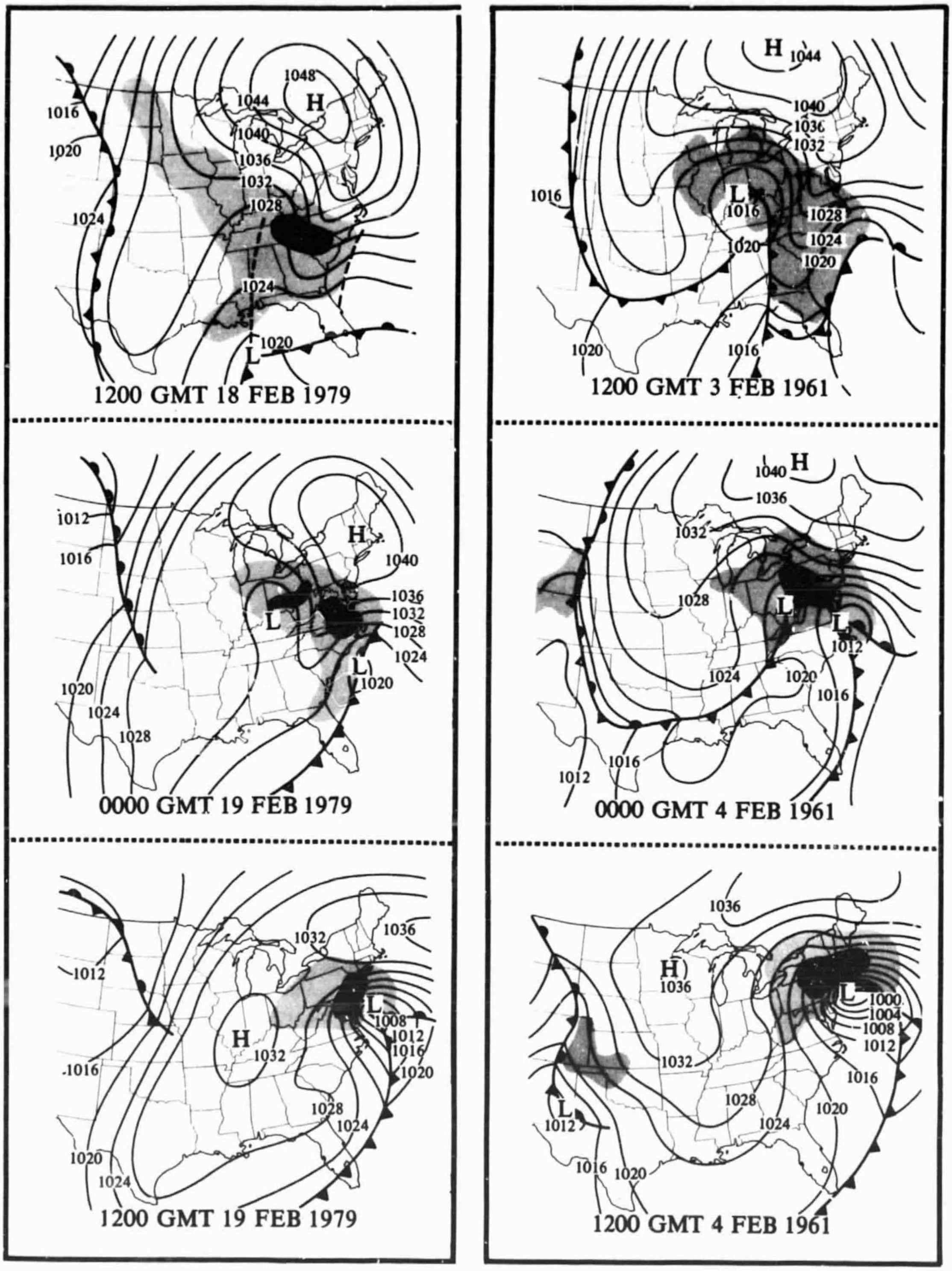


Fig. 4. Two similar examples of cyclone development at sea-level, 1200 GMT 18 February to 1200 GMT 19 February 1979 and 1200 GMT 3 February to 1200 GMT 4 February 1961. See Fig. 3 caption for details.

(Bosart, 1981; Uccellini et al., 1984, 1985). The February 1961 storm, which was responsible for record snow accumulations across central and southeastern New York, northern New Jersey, and parts of southern New England, is characterized as type "B." Despite the differences in classification, both storms share the following characteristics: (1) large cold anticyclones with record or near-record cold preceded each storm; (2) a wedge of high pressure extended southward along the East Coast deep into the southeastern United States; (3) the primary Ohio Valley surface low was weak and associated with mostly light precipitation; (4) coastal frontogenesis occurred off the Southeast United States coast; and (5) coastal low pressure centers deepened rapidly as they moved north-northeastward along the coastline before drifting to the east. Both cyclones also marked the end of two unprecedented periods of cold weather across the eastern United States (Foster and Leffler, 1979; Bosart, 1981; Ludlum, 1961).

b. Propagation Rates

To document the variations in movement of the individual surface low centers, average speeds (in m s^{-1}) were computed over 12 h periods during the 60 h period chosen to study each surface low center. The range of the average speeds for both primary and secondary low pressure centers is shown in Table 2. It appears that not only are there considerable case-to-case variations in speed, but there are also significant variations during the course of individual cases. The February 1961 and January 1964 snowstorms are two examples in which both primary and secondary low pressure centers moved at a relatively slow rate, averaging 9 to 12 m s^{-1}. In contrast, the

February 1967 surface low covered as much as 1200 km in 12 h, averaging 28 m s^{-1} as it propagated from the Gulf Coast to off the Middle Atlantic Coast. In eleven of the eighteen cases, there was a tendency of the surface low to reduce its average speed over a 12 h period by at least 5 m s^{-1} along the Northeast United States coast. The "stalling" of the sea-level low pressure center occurred most frequently from off the New Jersey coast to southern New England and its most obvious effect was to prolong the snowfall, especially in New England. The storms of March 1960, early and late February 1969, December 1969, and February 1978 are examples in which the surface low moved 250 km or less over a 12 h period, resulting in excessive snowfall amounts. The decrease in speed seems to occur as a manifestation of the occlusion process as surface and upper-level low centers become nearly colocated in the vertical [see Petterssen (1956) and Palmen and Newton (1969)].

c. Coastal Frontogenesis, Cold Air Damming and Canadian Anticyclones

Coastal frontogenesis along the East Coast of the United States (Bosart et al., 1972; Bosart, 1975; McCarthy, 1977; Marks and Austin, 1979) appears to have an important bearing on a large number of East Coast storms (Table 2). The coastal front is usually noted on NMC analyses as a stationary front imbedded within an inverted pressure trough near the coastline. Bosart et al. (1972) examined coastal frontogenesis off the New England coast and found that these shallow fronts tended to develop 6 to 12 h after a cold anticyclone became established to the north of the region. The coastal front typically develops near the coastline and is linked to ageostrophic deformation patterns, frictional effects, orography,

coastline shape and land-sea temperature contrasts. The fronts appear to act as channels for wave disturbances since they are the sites of enhanced coastal convergence, baroclinicity, warm advection, and surface vorticity. In a related paper, Bosart (1981) notes that the development of an intense coastal front and low-level baroclinic zone formed as cold, high pressure was modified through sensible and latent heat fluxes over a strong horizontal sea-surface temperature gradient. Recent modeling studies (Ballentine, 1980; Stauffer, 1984) have reproduced characteristics of the shallow coastal front and show that many factors, including large-scale troughs, surface friction, and sensible and latent heating, contribute to the highly ageostrophic deformation that leads to frontogenesis.

Coastal frontogenesis was observed in thirteen of the eighteen cases and occurred primarily from the Georgia to North Carolina coasts. These observations suggest that coastal frontogenesis is a significant component of the pre-cyclogenetic period and influences precipitation rates, type and distribution during cyclogenesis in many cases. As indicated in Table 2, coastal frontogenesis generally preceded the secondary development common to type "B" storms, but was absent in half of the type "A" cases. However, there are exceptions. The strong secondary development "B" cases of 9-10 February 1969 and 5-7 April 1982 were characterized by weak coastal frontogenesis as land-sea air temperature differences were rather small. In contrast, an intense coastal front occurred with the recent February 1983 snowstorm, which is classified as type "A." While coastal frontogenesis and cold air damming are frequently involved in these intense storms, they are not present in all cases, as shown in Table 2. For

example, the 6-7 February 1978 storm underwent explosive cyclogenesis without the apparent influences of damming or coastal frontogenesis.

Coastal fronts frequently form during the same period when cold air becomes "dammed" up against the Appalachian Mountains. Damming frequently occurs after a cold anticyclone or an associated ridge axis has become established over the northeastern United States and is viewed hydrostatically in the surface pressure field as a ridge of high pressure extending southward between the Appalachian Mountains and the coast, with the axis of the ridge reflecting the greatest depth of cold air in the layer (Baker, 1970; Richwien, 1980). Cold air damming remains inadequately forecast by operational numerical models (Richwien, 1980), possibly due to lack of proper vertical resolution in the model's planetary boundary layer and a poor representation of the topography. The inability of the operational models to adequately resolve damming results in temperature forecasts that are frequently too high, causes difficulties in delineating frozen versus liquid precipitation, and misrepresents the gradients in the lower-tropospheric thermal field, a crucial element for cyclogenesis.

Bosart et al. (1972), Richwien (1980), and Stauffer (1984) note that damming and its associated ageostrophic motions play a role in the formation of coastal fronts. A pattern of shearing deformation conducive to producing frontogenesis near the coast occurs with ageostrophic north to northeasterly flow between the Appalachians and the coast, and the east to southeasterly flow just east of the coastal front. Figs. 3b and 4 provide three examples of East Coast cyclones in which cold air damming is clearly apparent. Each case shows the distinctive high pressure wedge that can extend as far south as Florida. In each of the cases where coastal

frontogenesis occurred, damming was also present although its influence varied widely from case to case.

Damming usually occurs with the presence of an anticyclone to the north of the Middle Atlantic states. The position of the anticyclone with respect to the coastline and its interaction with the cyclone influences the evolution of the lower-tropospheric thermal field, the amount of moisture flux into the system, and thus is an important factor in determining the amount and the form the precipitation will take. Fig. 5 shows the locations of surface high pressure centers and their associated ridge axes for the eighteen cases at a time when the surface low pressure centers had reached the Atlantic Coast (as indicated by the darkened circles) where many were deepening rapidly (to be discussed shortly). For a majority of cases, the surface high pressure center or ridge axis was located from a region bounded in the west by Lake Superior and the southern tip of Hudson's Bay to the northeastern United States and extreme southeastern Canada as the surface low pressure center was developing along the East Coast. Many of these air masses were accompanied by unusually cold air. Damming is represented by the ridge axes that parallel the East Coast from New England to the Carolinas. In six cases, a surface high pressure center was located in a small region between the southern tip of Hudson's Bay and Maine. For the cases where the surface high pressure center or ridge axis was located north or northwest of New York, it appears that low-level air trajectories passing around the anticyclone into East Coast locations passed mainly over land. Thus, warming due to sensible heat fluxes from the Atlantic Ocean would be minimized, reducing the chances for snow to change to rain.

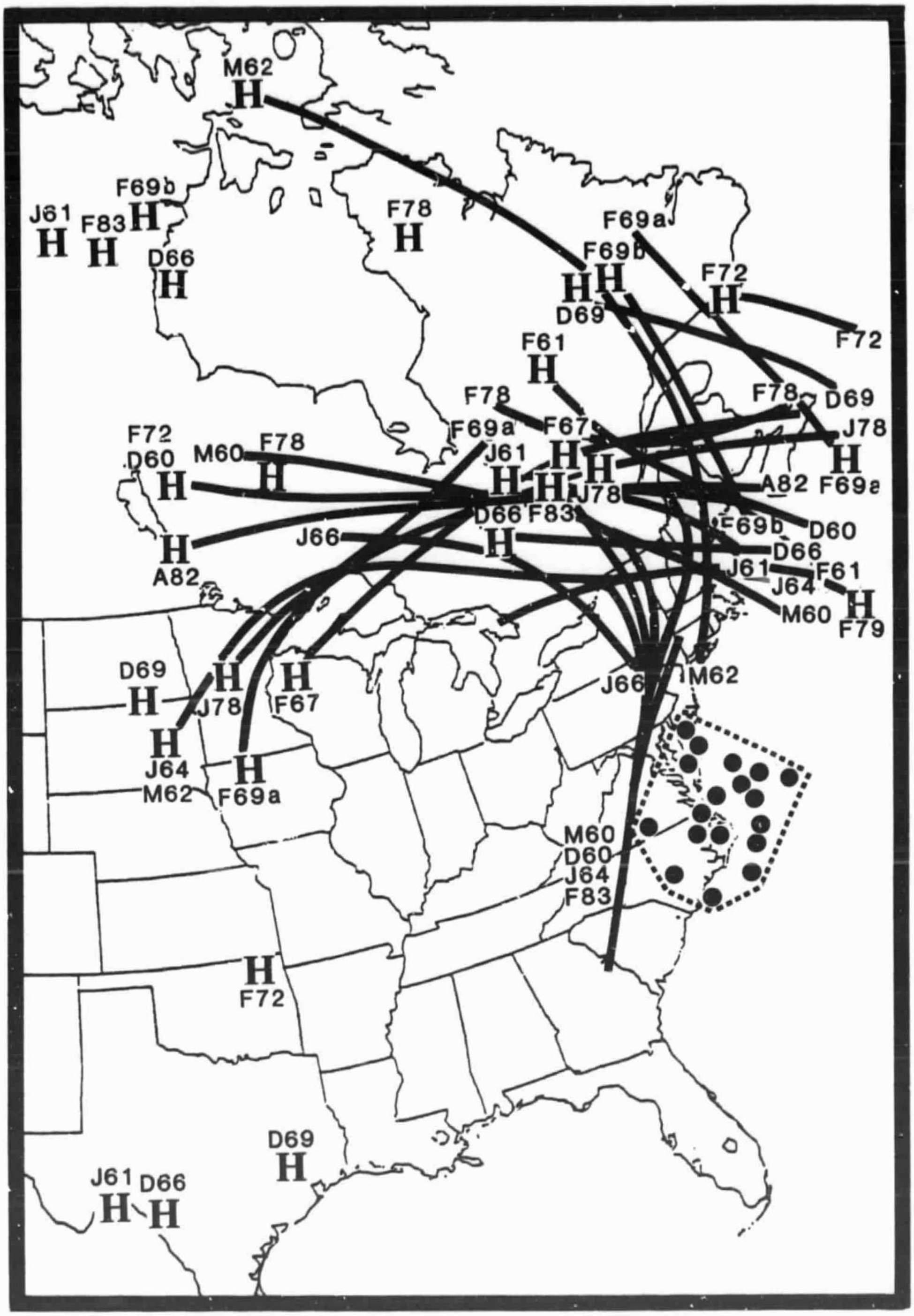


Fig. 5. Positions of surface high pressure centers (denoted by an H) and associated ridge axes (solid lines) at a standard 12-hourly synoptic reporting time when each of the eighteen surface low pressure systems was located off the Middle Atlantic coast (enclosed area; locations are marked by darkened circles).

The storms of March 1962, February 1969, December 1969, and February 1972 occurred in a regime in which the surface high or ridge axis was to the northeast of New England, setting up a longer fetch of air over the ocean than an anticyclone located further to the west. These snowstorms were not associated with the bitter cold that characterized the other storms. The December 1969 and February 1972 storms are two examples in which precipitation along the Northeast coastal regions began and ended as snow, but fell as rain in the interim. The early February 1969 snowstorm that buried New York City and eastern New England (see Fig. 1) was an unusual case in that no strong surface high preceded the storm. Thus, while the presence of a dome of unusually cold air was associated with many of these storms, and undoubtedly helped establish an intense low-level baroclinic zone, a few storms were not associated with such an air mass.

d. Deepening Rates

Deepening rates and central sea-level pressures were examined to provide measures of the intensities of the storms. Since many of the cyclones intensified over the data sparse ocean, NMC's analyses may misrepresent some of the actual sea-level pressures. Sanders and Gyakum (1980) studied rapidly deepening extratropical low pressure systems and characterized the most intense storms, or "bombs," as those that deepened at a rate of approximately 1 mb h^{-1} or more over 12 h periods. They noted that the East Coast of the United States and the adjacent Atlantic Ocean were deemed favorable locations for rapidly developing cyclones, especially over oceanic regions marked by large gradients of sea-surface temperature.

Central sea-level pressures of the eighteen storms are plotted (for types "A," "B," and "AB") at 3-hourly intervals during the 60 h period that each storm was studied (Figs. 6, 7 and 8). Table 3 summarizes the deepening aspects of all the storms, including the duration (h) and amount (in mb) of deepening ($\geq$-1 mb $(3\ h)^{-1}$) and rapid deepening ($\geq$-3 mb $(3\ h)^{-1}$) for primary and secondary sea-level low pressure centers. As shown in Figs. 6 through 8 and Table 3, sea-level deepening (defined by a decrease of pressure of at least -1 mb $(3\ h)^{-1}$ at the center of the low) occurred generally over a 24 to 48 h period for all eighteen storms. The amount of deepening ranged from 19 mb for the January 1978 storm to 52 mb for the March 1960 and January 1961 cases over the 60 h periods reviewed. However, given the observational constraints over the ocean, larger deepening rates are possible, but difficult to verify.

Type "A" storms, except for January 1978 and February 1983, deepened rapidly (exceeding -3 mb $(3\ h)^{-1}$) for 15 to 27 h during which the central pressures of the surface lows fell by 20 to 43 mb (Table 3, Fig. 6). For the eleven cases in which both a primary and secondary low pressure center (type "B" or "AB") was observed (Table 3, Figs. 7 and 8), the primary low pressure center usually did not deepen rapidly. Five of the eleven primary centers exhibited a period of rapid pressure falls (greater than -3 mb $(3\ h)^{-1}$), and only one case (April 1982) sustained it for more than 3 h. Once the secondary low pressure center had formed, the primary low coexisted with the secondary center anywhere from 6 to 27 h (for the type "B" cases) before being absorbed in the expanding circulation about the developing secondary cyclone (Table 3). All secondary low pressure centers underwent a 12 h or longer period of rapid deepening, with a longest period

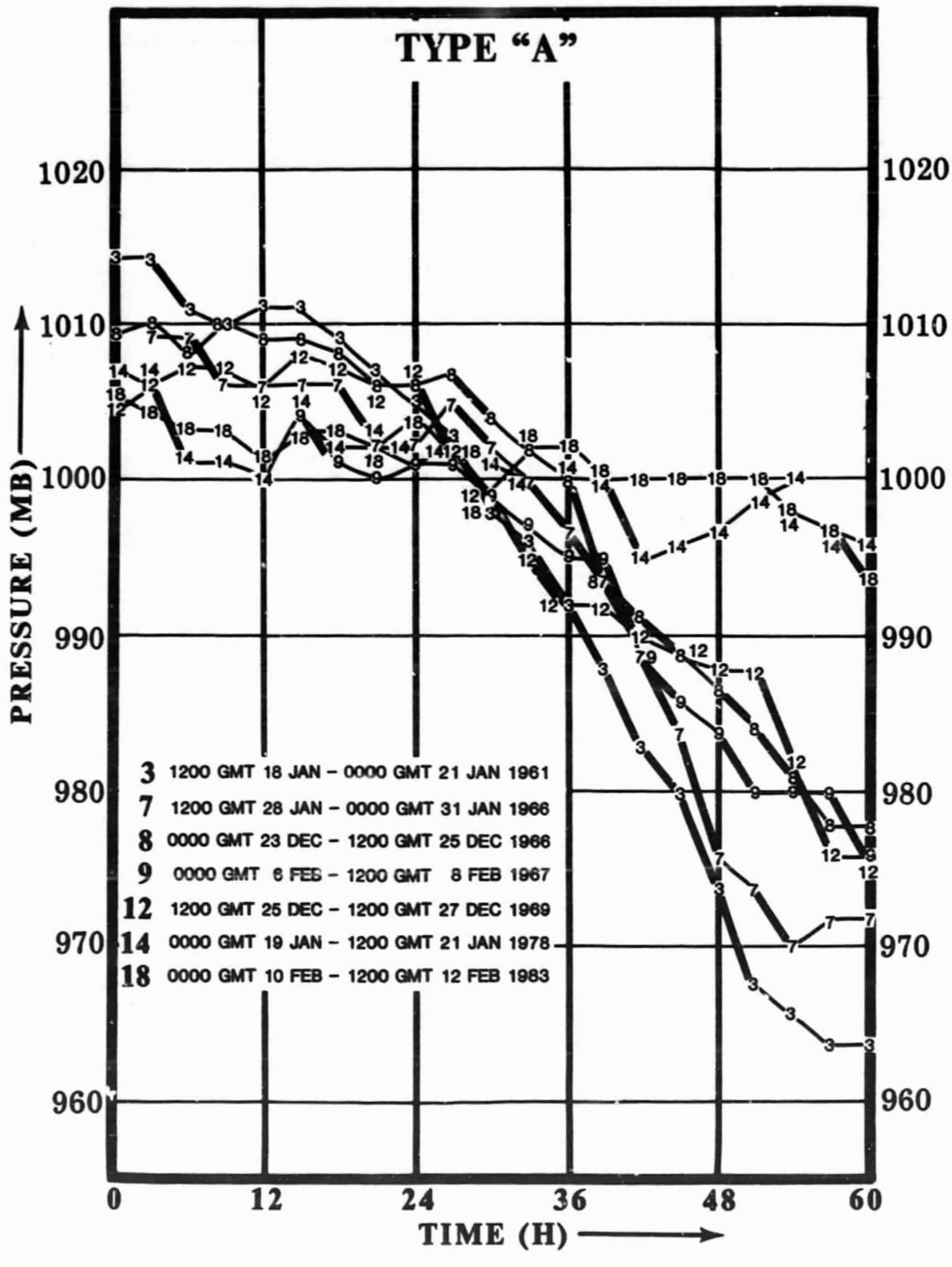


Fig. 6. Three-hourly traces of the central sea-level pressure of type "A" surface low pressure centers over the 60 h study period. Thick solid lines indicate rapid deepening (central sea-level pressures fell at -3 mb (3 h)$^{-1}$ or greater).

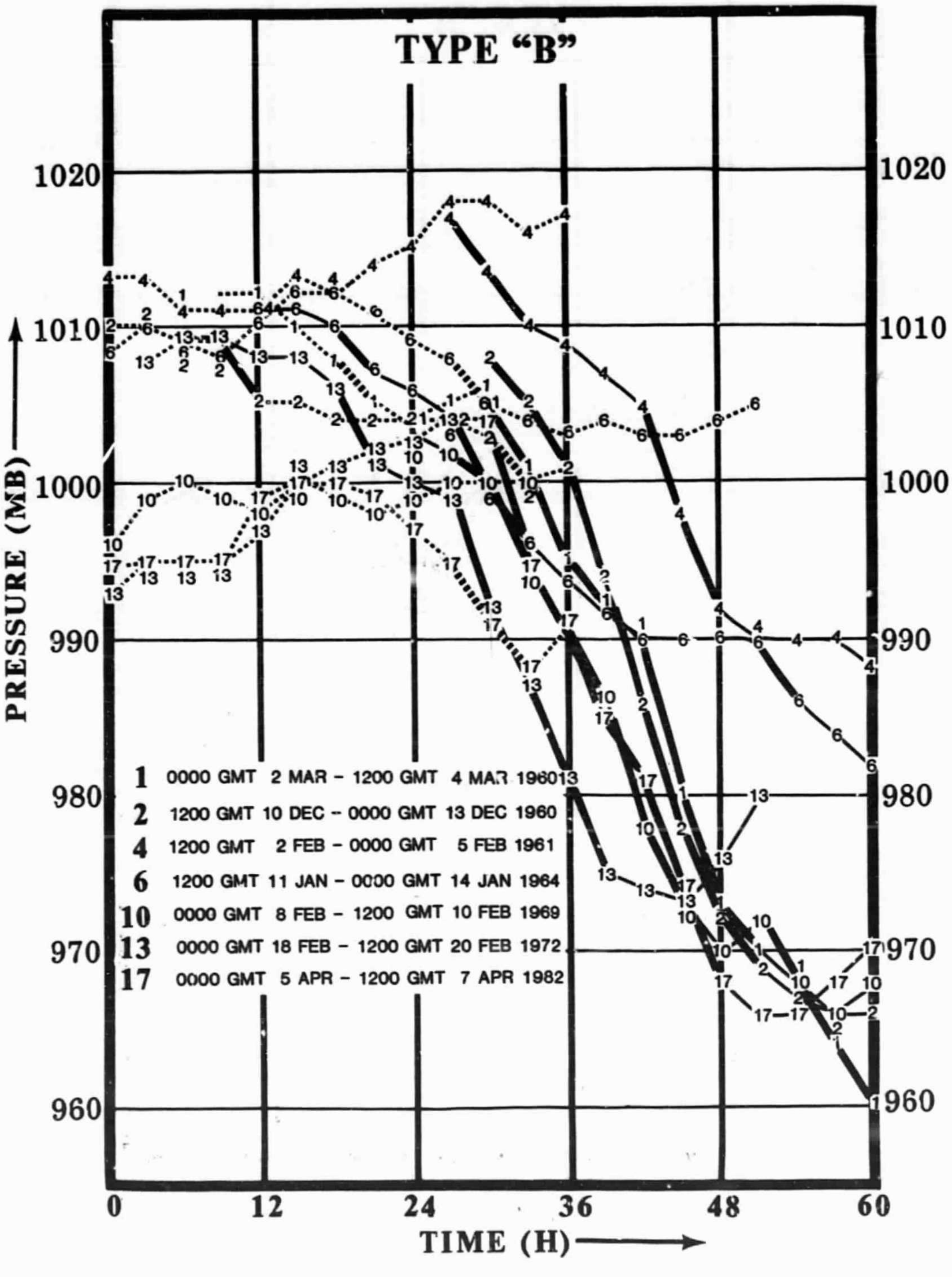


Fig. 7. Three-hourly traces of the central sea-level pressure of type "B" primary (dotted lines) and secondary (solid lines) low pressure centers over the 60 h study period. Thick dotted and solid lines indicate rapid deepening (central sea-level pressures fell at -3 mb $(3\ h)^{-1}$ or greater).

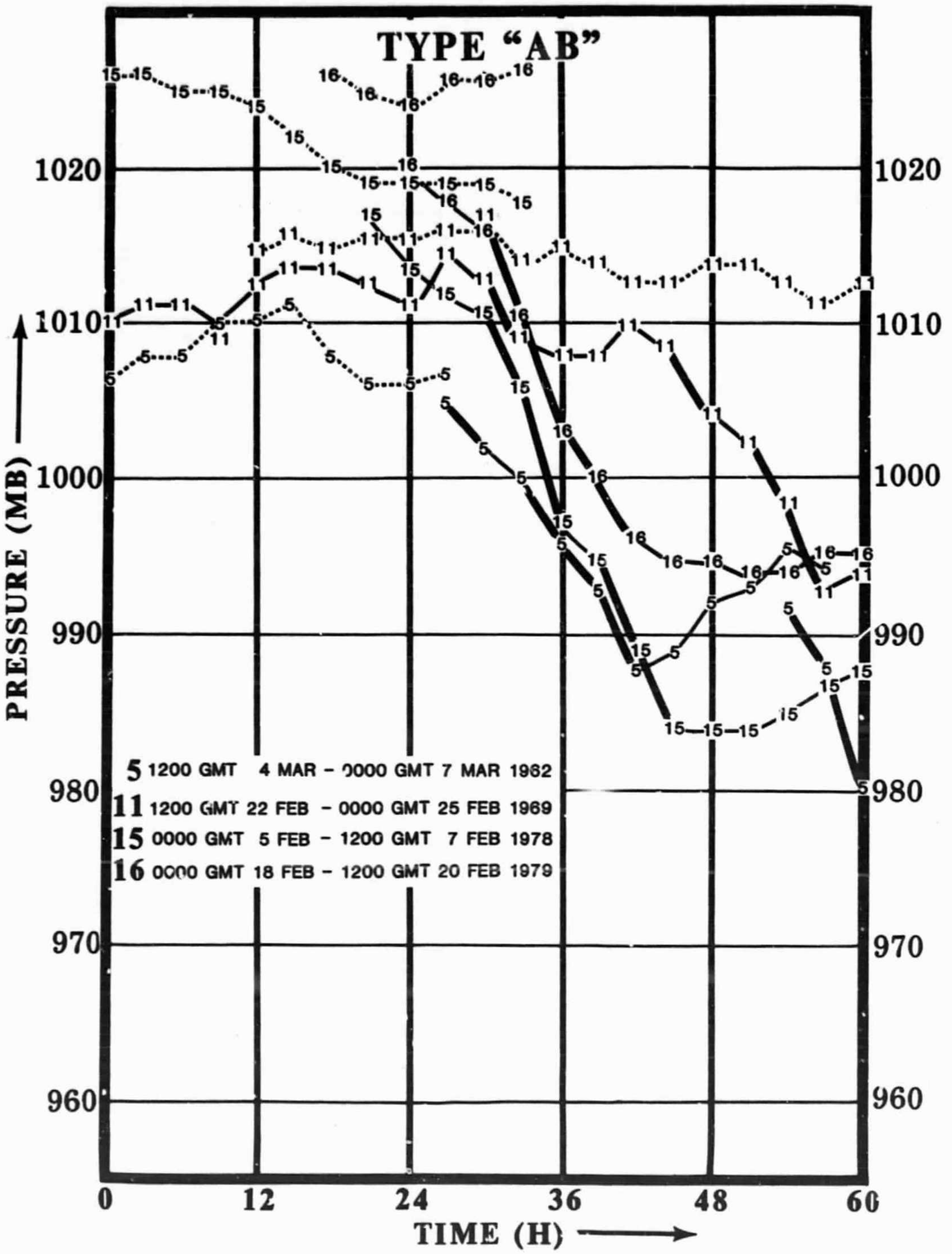


Fig. 8. Three-hourly traces of the central sea-level pressure of cases that could be categorized as either type "A" or "B" ("AB"). Traces are shown for both "primary" (dotted lines) and "secondary" (solid lines) low pressure centers over the 60 h study period. See Fig caption for details.

Table 3
Aspects of Sea-Level Deepening

Date	Duration (h) and [amount (mb)] of sea-level deepening ($\geq$-1 mb $(3\ h)^{-1}$)	Duration (h) and [amount (mb)] of rapid sea-level deepening ($\geq$-3 mb $(3\ h)^{-1}$) Type A	Duration (h) and [amount (mb)] of rapid sea-level deepening ($\geq$-3 mb $(3\ h)^{-1}$) Type B, AB Primary	Duration (h) and [amount (mb)] of rapid sea-level deepening ($\geq$-3 mb $(3\ h)^{-1}$) Type B, AB Secondary	Duration (h) of primary low once the secondary low formed (for type "B" cases only)	Lowest central sea-level pressure (mb)
M 60	42 (52)		3 (4)	24 (41)	15	960
D 60	39 (44)		3 (4)	21 (39)	6	966
J 61	48 (52)	24 (43)				964
F 61	33 (30)		----	12 (20)	9	988
M 62	27 (31)		3 (3)	18 (27)		980
J 64	45 (30)		3 (3)	12 (15)	24	982
J 66	33 (42)	27 (38)				970
D 66	42 (31)	18 (21)				978
F 67	30 (29)	15 (20)				976
F 69a	42 (39)		----	18 (32)	9	966
F 69b	27 (23)		----	12 (17)		993
D 69	36 (33)	18 (26)				976
F 72	33 (36)		----	15 (28)	27	973
J 78	30 (19)	6 (10)				995
F 78	39 (42)		----	15 (27)		984
F 79	27 (24)		----	12 (19)		995
A 82	33 (36)		6 (7)	18 (34)	6	966
F 83	30 (18)*	6 (6)				994

Table 3. Sea-level deepening aspects, including the duration (h) and amount (mb) of sea-level deepening ($\geq$-1 mb $(3\ h)^{-1}$) for all cyclones; the duration (h) and amount (mb) of rapid sea-level deepening ($\geq$-3 mb $(3\ h)^{-1}$) for type "A," type "B" and type "AB" primary and secondary low pressure centers; the duration (h) of the existence of a primary low pressure center once the secondary low pressure center had formed (for type "B" cases only); and the lowest central sea-level pressures (mb) of primary and secondary low pressure centers observed during the 60 h examination period.

*Deepening continued after study period ended.

of 21 h, in which the central pressures of these cyclones fell by 15 to 39 mb. In addition, the majority of storms seemed to undergo a clearly defined and continuous period of intensification prior to and during the period of very heavy snowfall. However, a few cases had a history of erratic intensification, especially those of January 1964 (Fig. 6) and December 1969 (Fig. 7). These two storms appeared to undergo two separate periods of rapid intensification that were separated by a 12 h period of little or no deepening.

All but two of the eighteen cases were associated with at least a 12 h period of rapid development, defined by central sea-level pressure falls of -3 mb $(3\text{ h})^{-1}$ or greater. In one of the two cases (February 1983), a 12 h period of rapid development did occur, but only as snow was ending across New England and the storm was moving well off the New England coast. Much of the heaviest snowfall from this storm fell at locations whose pressures were actually rising in response to the influence of gravity wave phenomena (Bosart and Sanders, 1985). The only case that did not intensify rapidly for 12 h or more occurred in January 1978, in which two brief and separated 3 h periods of rapid deepening were observed.

The paths of the low pressure systems during those periods when central sea-level pressures fell by -3 mb $(3\text{ h})^{-1}$ or more are shown in Fig. 9. The clustered paths of the rapidly developing storms from the North Carolina to southeastern New England coastlines indicate that processes along the East Coast probably contributed to the rapid development phase of cyclogenesis. Furthermore, the southwest to northeast orientation of the paths for these cases indicates that the storm centers

PATHS OF SURFACE LOW PRESSURE CENTERS DURING RAPID INTENSIFICATION

(> -3 MB (3 H)$^{-1}$)

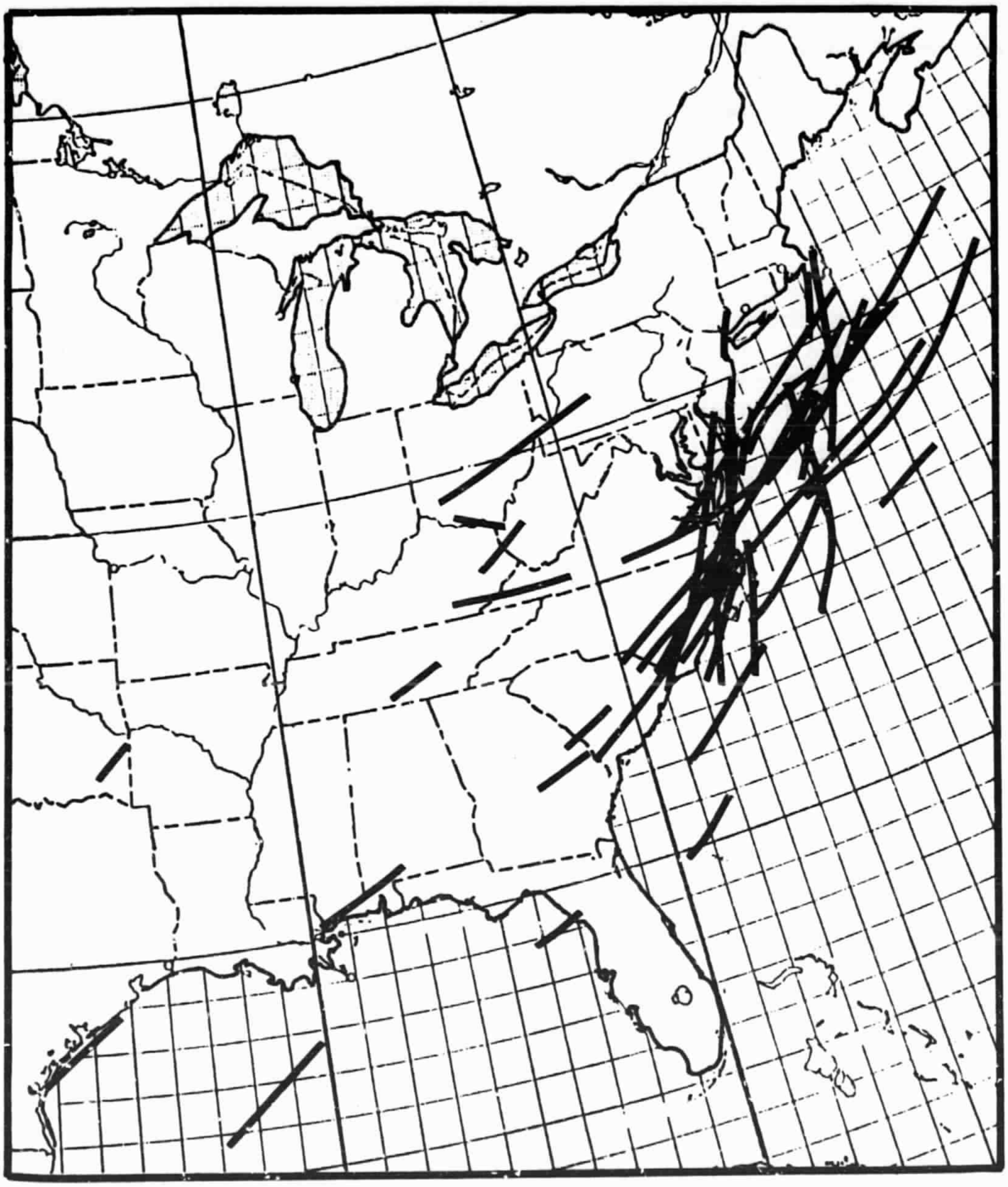


Fig. 9. Paths for all low pressure centers during rapid intensification (sea-level central pressure fell at -3 mb (3 h)$^{-1}$ or greater).

remained far enough from the coast to ensure that precipitation fell as snow, rather than rain, in the coastal urban centers.

Many of the cyclones attained very low minimum sea-level pressures, reaching 980 mb or less in eleven cases during the 60 h period they were examined (see Table 3). The deepest cyclone of the sample is the March 1960 snowstorm, whose central sea-level pressure bottomed out close to 960 mb just off southeastern New England early on 4 March. However, not all of the storms were characterized by exceptionally low central pressures. The January 1978, February 1979, and February 1983 storms all produced excessive snow amounts while each storm's central pressure remained near or above 1000 mb. These and other storms developed in a regime dominated by unusually high pressure. For example, the 6-7 February 1978 cyclone is generally acknowledged as one of the most severe winter storms to strike the Northeast this century. Yet, it deepened to only 984 mb at its most intense stage off Long Island late on 6 February, placing thirteenth out of eighteen cases in terms of lowest sea-level pressure. However, this storm developed only one day after it was a mere 1025 mb trough line in the Ohio Valley separating a 1055 mb anticyclone over southern Canada from another 1035 mb anticyclone located over the northeastern United States.

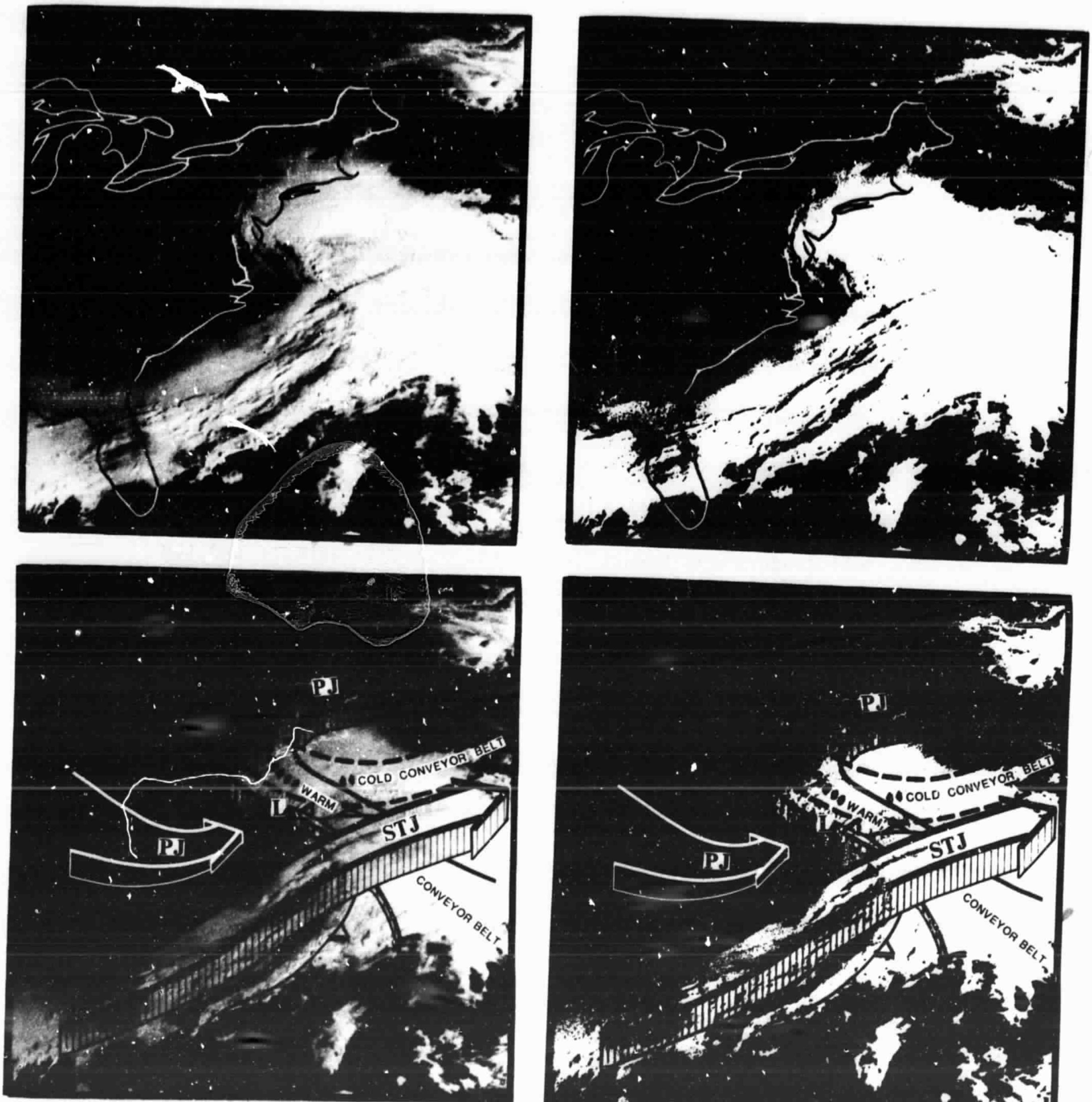
Many of the cyclones were accompanied by high winds, capable of producing deep drifts and reducing visibilities to near zero, which resulted from large pressure gradients between the low center and the anticyclone located generally to its north. The majority of cases seemed to develop the largest gradients in the direction of motion, or northeast of the surface low, accompanied by strong northeasterly winds. The

combination of an intensifying surface low and the decreasing distance between the low and downstream high, as the forward motion of the developing low becomes more rapid than the motion of the anticyclone, is frequently observed in these areas. The most intense pressure gradient and strongest winds are focused along the coast, where the effects of friction are minimized. The increased easterly flow north of the developing cyclone also enhances the moisture flux toward the coastal zone. In some instances, such as the storms of 29-31 January 1966 and the March 1888 "Blizzard of '88" (Kocin, 1983), the sea-level pressure gradients were largest to the west of the surface low and strongest winds occurred from a northwesterly, rather than a northeasterly, direction. While the minimum pressure of the February 1978 storm was not exceptionally low, its interaction with an intense anticyclone resulted in the formation of an intense sea-level pressure gradient immediately north and east of the low center, with a 10 mb difference between Nantucket and Boston, Ma. at 0000 GMT 7 February. This gradient and its associated winds were responsible for blizzard conditions along the New England coast, placing this storm in a class by itself in terms of paralyzing snowstorms.

e. Satellite Imagery

The introduction of continuous satellite imagery in the mid 1970's has enabled researchers to study cloud signatures of a multitude of phenomena, including mesoscale convective complexes (Maddox, 1980), flash flooding (Spayd and Scofield, 1983), extratropical cyclones (Weldon, 1979; Carlson, 1980; Scofield et al., 1982), and East Coast snowstorms (Parmenter, 1972).

Five of the eighteen storms studied here include 12 h sequences of infrared satellite imagery (see Part 2) that reveal some characteristic signatures of intense East Coast snowstorms. The most common characteristic of the five cases is the comma-shaped cloud, as depicted by the visible satellite GOES-East image of the February 1979 "Presidents' Day" storm during a period of rapid intensification (Fig. 10). Heaviest snowfall at this time is located from Washington, D.C. to New York City, along the southwestern edge of the cold cloud tops that make up the comma "head"; also noted by Carlson (1980), Scofield et al. (1982), Uccellini et al. (1984), and Sanders and Bosart (1985a,b). In this example and the intense storm of February 1978, explosive cyclogenesis was accompanied by the rapid expansion and vertical growth (colder cloud tops) of the comma head. The growth of the comma head appears to be a result of ascent associated with two airstreams described as warm and cold "conveyor belts" (Carlson, 1980). From Fig. 10, there is no way to distinguish the two airstreams. As discussed by Carlson, the warm conveyor belt originates in the warm sector of the cyclone, ascends toward the north, reaches saturation near or north of the warm front, and joins the upper-level westerly flow poleward of the surface low center. The cold conveyor belt forms on the cold side of the warm front toward the region north of the surface low center and passes beneath the warm conveyor belt. Saturation in the cold conveyor belt occurs from precipitation falling into it from the warm conveyor belt and by ascent within the exit region of a polar jet streak. The warm conveyor belt and the comma head frequently have a distinct anticyclonically curved western and northern edge that align with the axis of the upper-level jet stream, as shown by Anderson et al. (1974),



1330 GMT 19 FEBRUARY 1979 **1330 GMT 19 FEBRUARY 1979**

Fig. 10. (Top) GOES-East visible satellite image of the "Presidents' Day" storm at 1330 GMT 19 February 1979. (Bottom) Schematic of air flow associated with the cyclone (after Carlson, 1980), including surface frontal and low positions, axes of subtropical and polar jets, and postulated locations of warm and cold conveyor belts.

Weldon (1979), Carlson (1980), and Thepenier and Cruette (1981). The comma tail, in this case, was associated with clouds near and along the surface cold front. An intrusion of dry air associated with a polar jet streak frequently aids in the production of the comma-shaped cloud as it gives the appearance of pinching off the comma head from the tail (e.g., see the 1830 GMT 19 February 1979 image in Fig. 11). Recent evidence by Carr and Mallard (1985) indicates that the dry intrusion in developing cyclones may actually be ascending near the cyclone center, but the air is so dry that cloud formation is inhibited. In general, all cases exhibited a more pronounced comma shape during cyclogenesis as the comma head expanded at an angle nearly normal to the orientation of the comma tail.

Another signature of these storms is the development of an "eye"-like cloud-free zone near the center of the cyclones once they have passed out over the Atlantic Ocean [also described by Sanders and Gyakum (1980), Bosart (1981) and Uccellini et al. (1985)], similar to that observed in tropical storms. Four of the five most recent cases with continuous satellite data exhibited this feature (Fig. 11), typically towards the end of the rapid deepening stage. The clouds that surround the cloud-free center are typically fairly shallow. Thus, they do not resemble the deep convection that usually surrounds the centers of mature tropical cyclones. Clouds may extend to fairly high levels well north of the cloud-free centers within ascending air (as part of the warm and cold conveyor belts shown in Fig. 10) while shallow cloudiness south of the clear zone is likely associated with subsiding air along the axis of the polar jet (Fig. 10), inhibiting the vertical growth of the clouds. Anthes and Keyser (1979) simulated an "eye-like" structure in an extratropical cyclone over

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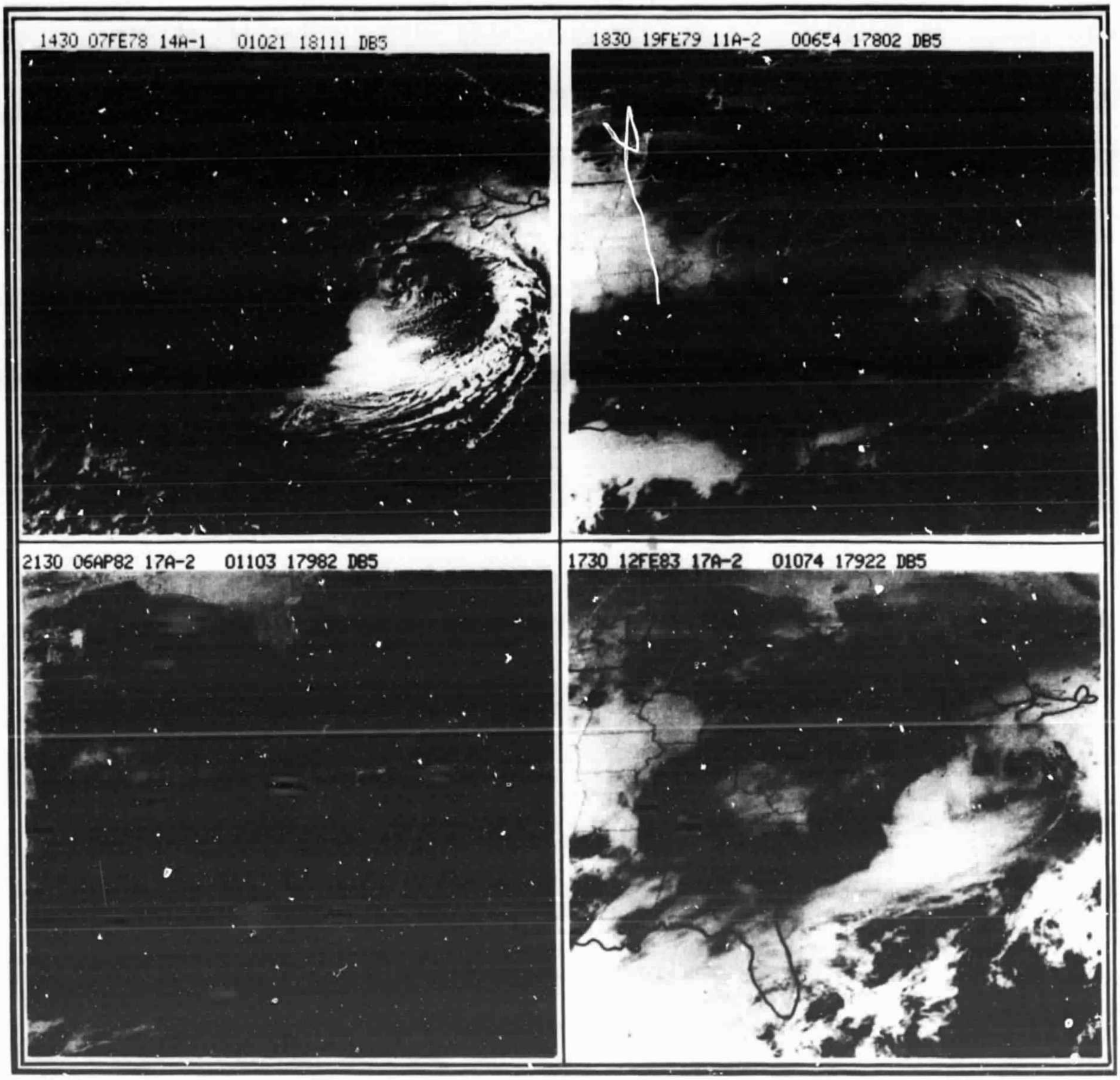

Fig. 11. Visitle GOES-East satellite imagery depicting "eye"-like cloud-free zone over the Atlantic Ocean in four recent major East Coast snowstorms; (a) 1430 GMT 7 February 1978; (b) 1830 GMT 19 February 1979; (c) 2130 GMT 6 April 1982; and (d) 1730 GMT 12 February 1983.

the Ohio Valley when they omitted friction in their numerical simulation, indicating that decreased friction over the ocean might aid in the development of this feature. The January 1978 storm was the only cyclone out of five without an "eye," but this cyclone differed from the others in that it did not maintain a significant period of rapid intensification.

3. OVERVIEW OF UPPER-LEVEL CHARACTERISTICS

Analyses of surface weather features indicate that the evolution of the surface cyclone can be categorized into two types. However, the upper levels portray a more complicated structure and evolution in terms of patterns of geopotential and wind fields which are composed of troughs and ridges of varying wavelength, amplitude and configuration. Multiple jet streak systems are imbedded within the trough/ridge systems and also evolve in ways that are difficult to categorize. Some of these jet streak systems amplified prior to the development of the surface low, others amplified during surface cyclogenesis while others did not amplify at all.

Upper-level features associated with the development of the eighteen East Coast snowstorms are listed in Tables 4 through 6. Table 4 summarizes the type, amplitude, and wavelength of the 500 mb trough/ridge systems associated with the storms, and also indicates whether diffluence, negatively tilted trough axes, and a separate trough/ridge system across eastern Canada and Greenland accompanied these cases. The characteristics of middle- and upper-tropospheric jet streams and imbedded jet streaks are summarized in Table 5, which highlights temporal and spatial changes in the magnitudes of wind speeds associated with jet streaks. Table 6 summarizes 850 mb features.

a. 500 mb Troughs and Ridges

The most conspicuous upper-level characteristic of every cyclone in this sample is the presence of a well-defined trough/jet streak system in the middle and upper troposphere, a situation which maximizes upper-level

divergence and associated positive vorticity advections as discussed by Palmen and Newton (1969), among many others. This feature was always observed during the two- to three-day period that preceded the formation or propagation of the surface low along the Atlantic coast. To provide a representation of the movement of the trough systems, centers of maximum geostrophic vorticity, inferred from the curvature and shears of the 500 mb geopotential height contours, were plotted at 12 h intervals, and their paths are shown in Fig. 12, grouped according to Miller's (1946) surface classification scheme. Many of the vorticity maxima are observed to move eastward or southeastward across the center of the country, and then northeastward up the Eastern Seaboard. The majority of vorticity maxima pass across the lower Ohio Valley and northern Gulf Coast states to the Virginia-North Carolina coast and then out over the Atlantic Ocean. While the southward displacement of vorticity maxima (indicative of "digging" troughs) was prominent in the early stages of cyclogenesis, heavy snowfall along the East Coast occurred only when vorticity maxima were propagating to the east or northeast. Also note that multiple vorticity centers for at least seven cases (e.g., December 1960 and January 1966) reveal that some upper-level trough systems involved the merging of two or more pre-existing troughs.

a1. Trough Evolution

Using the 500 mb analyses, three categories define transitions in the configurations of the upper-level troughs during the period when the surface low moved to or intensified along the East Coast (Table 4). The first category consists of traveling "open wave" troughs that evolve into a closed vortex while sea-level cyclogenesis is in progress. The closed

TRACKS OF 500 MB VORTICITY MAXIMA

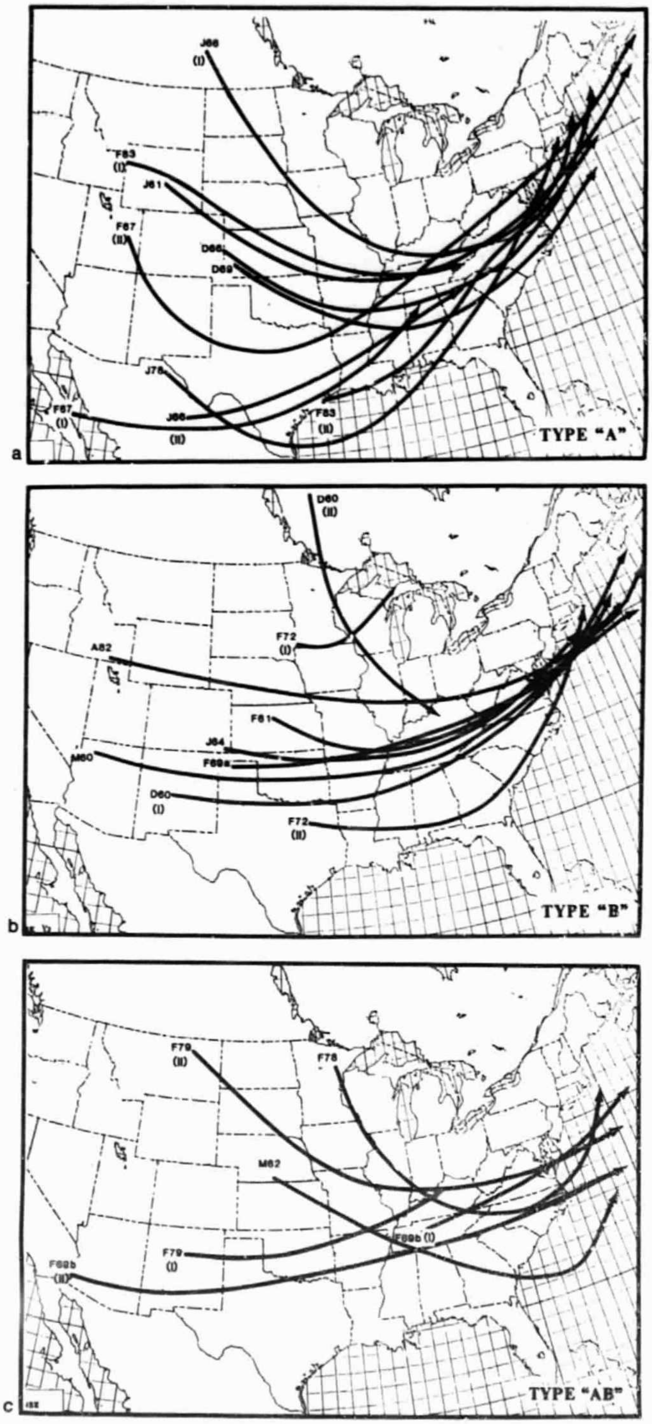

Fig. 12. Tracks of inferred 500 mb geostrophic vorticity maxima for all storms [grouped according to Miller's (1946) criteria]. Multiple maxima are denoted by (I) or (II).

Table 4
Summary of 500 mb Trough Characteristics

Date	500 mb Trough Category Open to Closed (O→C); Open (O); Closed (C)	Maximum Number of Closed 500 mb Contours	Maximum Amplitude Measure (° latitude) (Maximum Displacement of Height Contours) 1 Upstream Ridge and Trough; 2 Trough and Downstream Ridge: 1	2	Amplitude Increase (>5°) Between Upstream Ridge and Trough Prior to time T	Changes in Amplitude (° latitude) between Trough and Downstream Ridge during Rapid Cyclogenesis	Shortening Half-Wavelengths between trough and Downstream Ridge* *P=Prior to Rapid Cyclogenesis, D=During Rapid Cyclogenesis
M 1960	O→C	3	16(T)	12(T+12)		+10°	X (P)
D 1960	Other	[illegible]	28(T)	15(T+24)		+14°	X (P)
J 1961	O→C	1	35(T+12)	15(T+24)	X	+14°	X (P,D)
F 1961	C	3	18(T)	9(T+12)		+5°	X (D)
M 1962	C	5	24(T+12)	14(T+12)	X	+5°	X (D)
J 1964	[illegible]	2	26(T)	15(T)		+5°	X (P,D)
J 1966	Other	6	23(T+12)	28(T+24)	X	+15°	X (P,D)
D 1966	O→C	4	29(T)	22(T+24)	X	+12°	X (D)
F 1967	O	0	30(T)	10(T+12)		+8°	X (D)
F 1969a	O→C	4	20(T)	19(T+12)		+11°	X (D)
F 1969b	O→C	1	14(T+24)	13(T+24)	X	+6°	X (?)
D 1969	O→C	1	18(T)	24(T+24)		+16°	X (D)
F 1972	O→C	2	27(T+12)	23(T+24)	X	+12°**	X (D)
J 1978	O	0	15(T-12)	16(T)		+5°	X (-)
F 1978	C	5	41(T+12)	16(T+12)	X	+12°	X (D)
F 1979	O	0	12(T)	7(T+12)	X	0	X (P,D)
A 1982	O→C	2	20(T+12)	11(T+24)	X	+6°	X (P)
F 1983	O	0-1	16(T)	9(T+24)		+5°	X (D)

** = While cyclone was moving up the East Coast (no rapid deepening)

Table 4. Summary of 500 mb trough evolution, amplitude and wavelength, including a trough categorization (open-wave troughs that evolve into closed-contour vortices, O→C; open-wave troughs that do not evolve into a vortex, O; and closed-contour troughs prior to and during cyclogenesis, C); the maximum number of closed height contours (at 60 m intervals) for troughs that develop a closed circulation at 500 mb; a measure of the maximum amplitude of the trough and flanking ridges [the maximum latitudinal displacement of height contours (in degrees) and time of occurrence (with respect to time T; see text)]; cases in which the amplitude (measured as the difference in degrees latitude along selected height contours between trough and ridge axes) between the trough and upstream ridge increased by 5° or more prior to rapid sea-level development; a measure of the increase of amplitude (in degrees latitude) of the trough and downstream ridge during rapid sea-level development; and cases in which the half-wavelength between the trough and downstream ridge decreases, and whether it decreased most prior to (P) or during (D) rapid cyclogenesis.

Table 4 (continued)
Summary of 500 mb Trough Characteristics

	Diffluence	Negative Tilt	Eastern Canadian Trough	Greenland NE Canadian Ridge
M 1960	X	X	X	X
D 1960	X	X?	X	
J 1961	X	X	X	
F 1961	X	X	X	
M 1962	X	X	X	X
J 1964	X		X	X
J 1966	X	X	X	X
D 1966	X	X	X	X
F 1967		X	X	
F 1969a	X	X	X	
F 1969b	X?	X	X	
D 1969	X	X	X?	
F 1972	X	X	X?	
J 1978	X?	X	X	
F 1978	X	X	X	X
F 1979	X	X	X	
A 1982	X	X	X	X
F 1983	X	X	X	X

Table 4 (cont'd). Summary of 500 mb height field characteristics, including cases exhibiting diffluence, a negatively tilted trough axis (trough axis rotated to a northwest to southeast orientation), a trough across eastern Canada, and a ridge across Greenland.

vortex is defined by at least one enclosed height contour (where the contours are spaced at 60 m intervals). The second category involves open wave troughs that do not develop a closed center at 500 mb prior to or during cyclogenesis along the East Coast. The third category consists of trough systems that maintain a closed 500 mb circulation prior to and during cyclogenesis on the East Coast.

The transition of an open wave system to a closed-center vortex at 500 mb was the most commonly observed category of trough systems in the sample. Eight of the eighteen cases clearly demonstrated this trend, which occurred for type "A," "B," and indeterminate cases of sea-level development (type "AB"). The transition from the open wave to a closed vortex was usually associated with the increased amplification of the trough/ridge system, decreased wavelength between the trough and downstream ridge axes, the development of a "negative tilt," where the trough axis extends from the northwest to the southeast, and the formation of diffluence immediately downwind of the trough axis. All of these cases were associated with rapidly deepening surface low pressure centers that maintained deepening rates of at least -1 mb $(h)^{-1}$ for 12 h or more (Table 3). In addition, the 500 mb vortices formed at least 6 to 12 h after the onset of rapid sea-level development, with the height at the center of the vortex decreasing by 60 to 120 m during the first 12 h following its formation. The number of closed contours (drawn at 60 m intervals) ranged from one to four for all of these systems (see Table 4). Since the upper-level vortex formed after the commencement of rapid sea-level intensification, the generation of a cut-off low at 500 mb may be viewed as the upward growth of the rapidly-developing cyclonic circulation

as the system occluded. Finally, the sea-level and upper-level low centers became nearly colocated in the vertical, a reflection of the occlusion stage as rapid sea-level development was diminishing.

An example of a system with these characteristics is shown in Fig. 13 which displays a transition from an open wave to a closed center and negatively tilted vortex at 850 mb for the April 1982 spring snowstorm. At 0000 GMT 5 April, a short wave trough at 500 mb over the northern Rocky Mountain states was associated with dual 994 mb surface low pressure centers moving off the eastern slopes of the Rocky Mountains with precipitation scattered across the Rockies and western Plains states. By 0000 GMT 6 April, the surface low pressure centers had consolidated into one over Oklahoma that then moved to the Ohio Valley associated with an expanding area of precipitation to its north and east. At the same time, the 500 mb trough configuration had changed character in several respects. First, a definite height minimum had formed and was located across western Wisconsin. In addition, the amplitude of the trough increased, the half-wavelength between the trough and downstream ridge decreased, the height gradients at the base of the trough increased significantly, and diffluence downwind of the trough axis became quite noticeable. In the following 12 h, explosive secondary cyclogenesis commenced along the Middle Atlantic coast and by 0000 GMT 7 April, a closed circulation formed at 500 mb, its central height deepening 180 m in 24 h, and the axes of the trough rotated in a counterclockwise sense and developed a negative northwest to southeast tilt. The surface low, which deepened rapidly to 968 mb, intensified beneath the diffluent portion of the trough, immediately downwind of the trough axis.

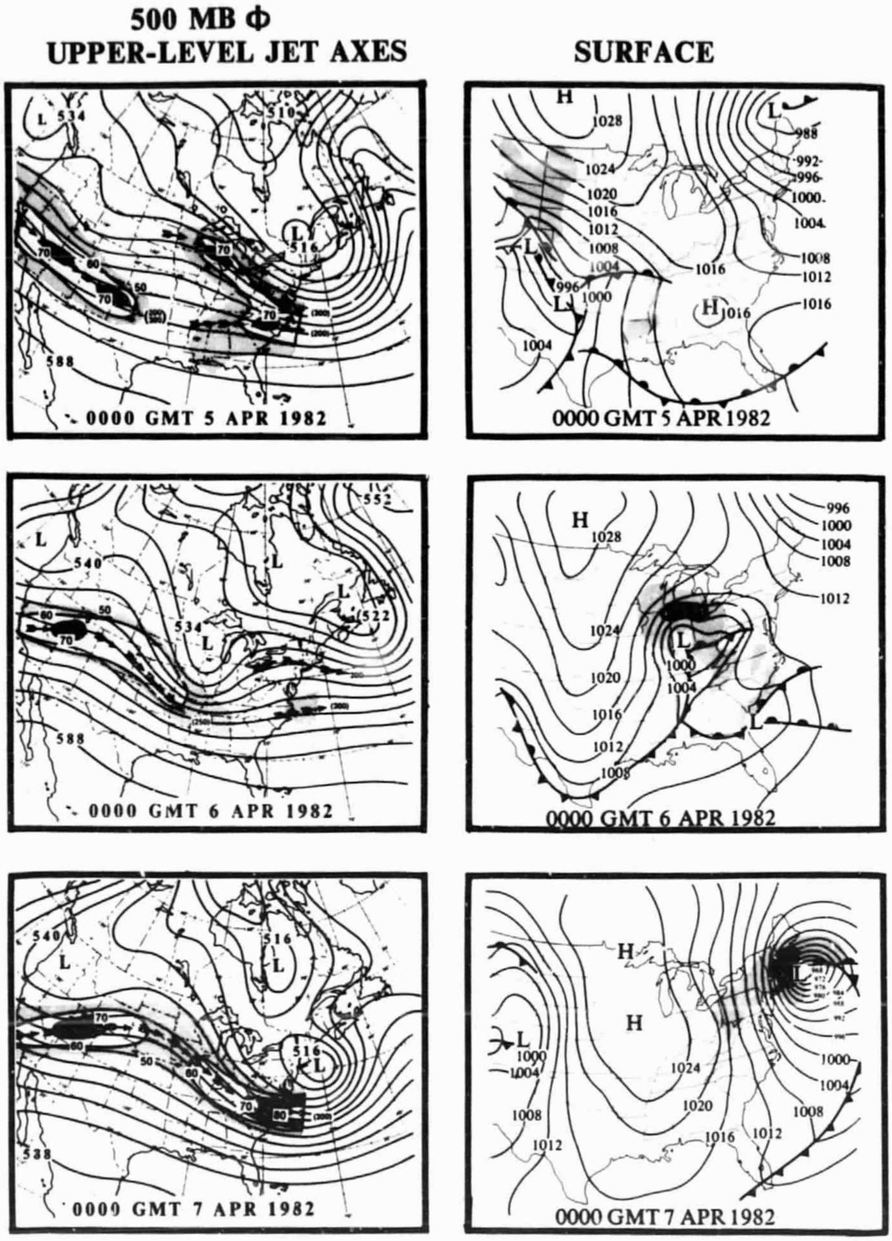

Fig. 13. Illustration of a 500 mb "open-wave" trough that evolves into a closed-center vortex, and corresponding surface analyses for 0000 GMT 5, 6 and 7 April 1982. (Left) 500 mb geopotential height contours (solid, 516 = 5160 m; contours are spaced at 60 m increments); axes of upper-level jet streams and isotachs of maximum wind speeds (derived from 500, 300, 250 and 200 mb charts; values in parentheses mark level of maximum wind speeds; shading represents wind speeds of 50 to 60 m s^{-1}; heavy shading represents wind speeds of 70 to 80 m s^{-1}. (Right) surface analyses. See Fig. 3 caption for details.

In four of the eighteen cases, an open wave trough did not evolve into a vortex at 500 mb during the 60 h period studied (Table 4). The four cases include the February 1967, January 1978, February 1979 and February 1983 storms. With the exception of the February 1979 "Presidents' Day" storm, whose surface characteristics suggest either type "A" or "B" development, the three other trough systems were associated with type "A" sea-level development. In general, this category is made up of storms that did not deepen as greatly as those in the previous category. However, rapid sea-level development occurred in February 1979 and February 1983 once the storms moved out over the Atlantic Ocean. In spite of relatively minimal upper-level amplification, these cyclones produced as much snowfall, over as large an area as the previous class of trough systems.

An example of an open wave trough system is provided by the recent February 1983 storm (Fig. 14). At 1200 GMT 10 February, a short wave trough was located along the Gulf Coast and a less distinct trough was situated further upstream across the Central Plains states. Dual surface low pressure centers were located along the Gulf Coast and associated with a region of light to moderate rainfall. By 1200 GMT 11 February, only one surface low pressure center had moved from the Gulf Coast up the Southeast Coast to a position near southeastern North Carolina. This low was not deepening, but was associated with a large region of moderate to heavy snow and rainfall throughout the Middle Atlantic states. At 500 mb, no amplifying trough was observed. The two troughs evident 24 h earlier had combined into one broad trough with a height minimum stretched across Kentucky. Heights at the center of the trough fell only 60 m during the previous 24 h, and its amplitude and half-wavelength between trough and

500 MB Φ
UPPER-LEVEL JET AXES **SURFACE**

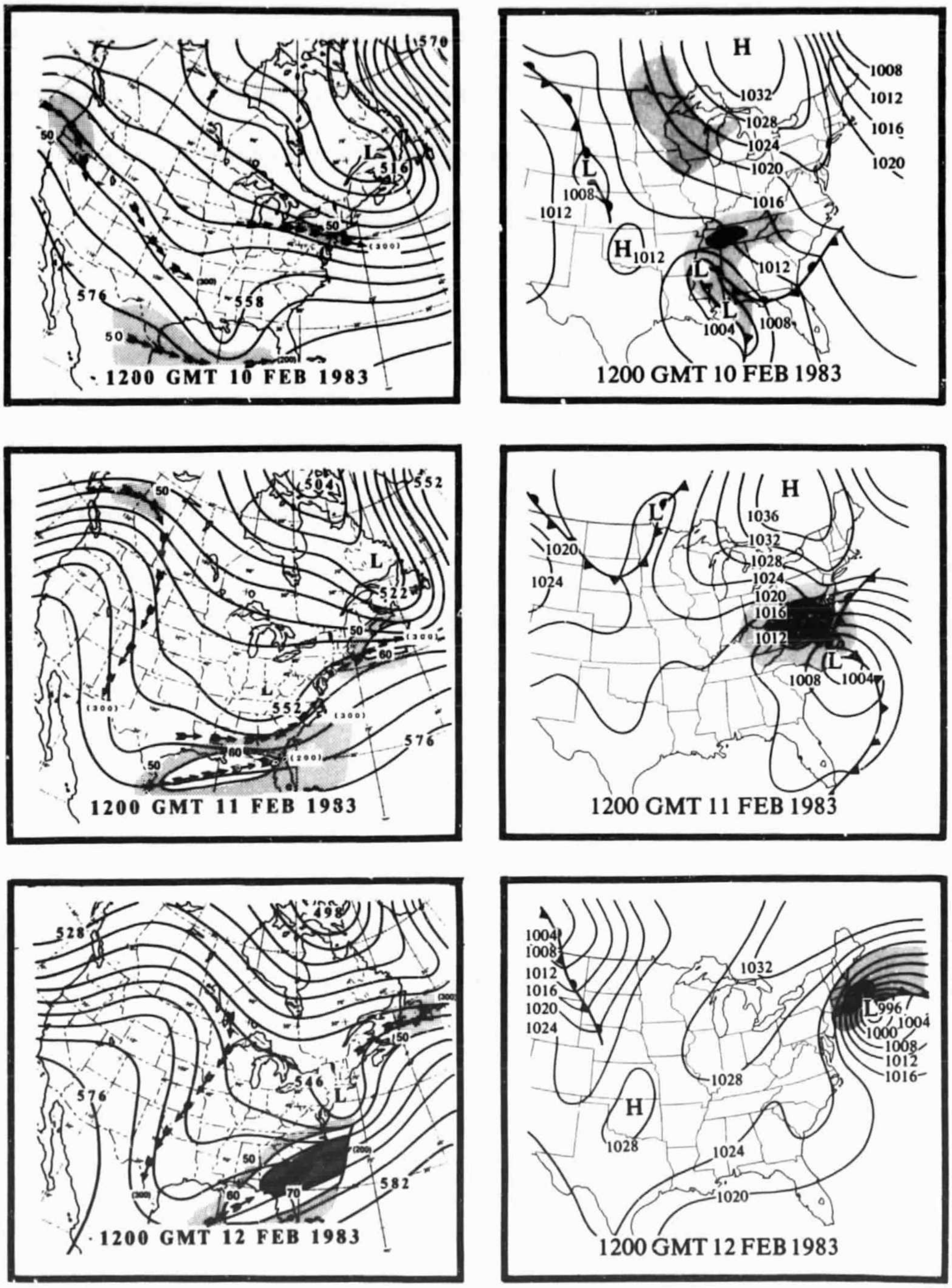

Fig. 14. Illustration of 500 mb trough that does not evolve into a closed-center vortex, and corresponding surface analyses for 1200 GMT 10, 11 and 12 February 1983. See Fig. 13 caption for details.

downstream ridge changed little. In the following 24 h, heavy snows moved through much of the coastal Northeast. By 1200 GMT 12 February, the surface low moved off the southeastern New England coast and had just begun to deepen rapidly as heavy snows were ending across eastern New England. At 500 mb, there is still no closed low, but heights in the center of the trough continued to fall. There are indications that the trough was now showing signs of amplification with a slight increase of amplitude, decrease of wavelength, increased geopotential gradients, and enhanced diffluence. Within 12 h, a deepening 500 mb closed center had formed in conjunction with a rapidly deepening surface system. However, the deepening occurred after the storm had departed the East Coast.

Four additional storms formed in association with upper-level trough systems that contained a closed center at 500 mb prior to a period of rapid sea-level deepening along the East Coast (Table 4). The four cases include the storms of February 1961, March 1962, January 1964, and February 1978. Each of these cases was associated with secondary sea-level development (type "B"), although February 1978 can be classified as either "A" or "B." Three of the four cut-off troughs (March 1962, January 1964 and February 1978) were massive systems that covered a substantial portion of the nation as they drifted slowly eastward, maintaining two to as many as six closed contours at 500 mb (Table 4). All four cases moved slowly relative to the other cases studied.

To illustrate the type of system characterized by a closed 500 mb vortex prior to cyclogenesis, a sequence of upper-level and surface charts for the January 1964 snowstorm is presented in Fig. 15. At 0000 GMT 12 January, a massive trough was located in the central United States with

500 MB Φ
UPPER-LEVEL JET AXES **SURFACE**

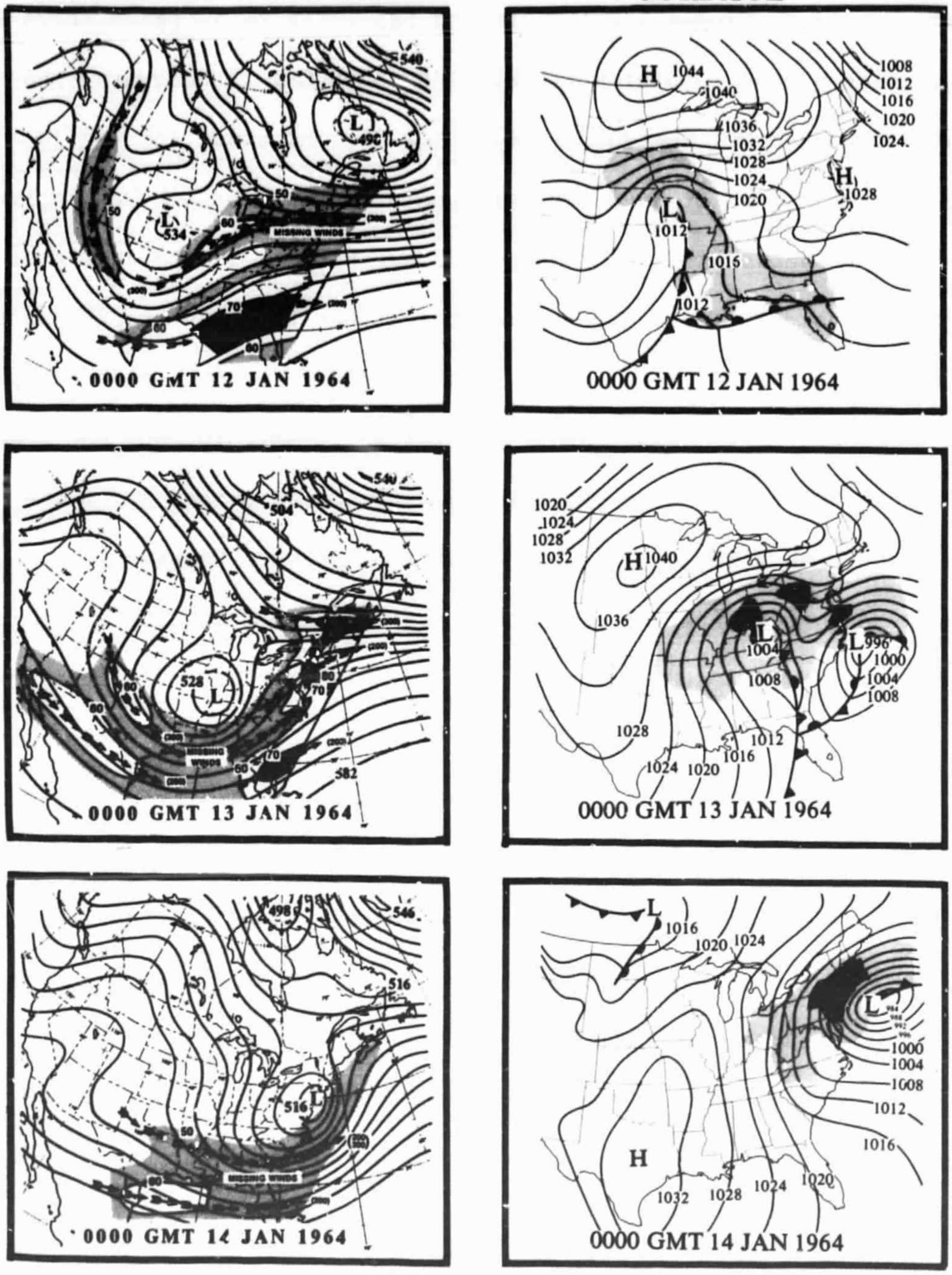

Fig. 15. Illustration of a closed-center 500 mb vortex prior to and during East Coast cyclogenesis, and corresponding surface analyses for 0000 GMT 12, 13 and 14 January 1964. See Fig. 13 caption for details.

a closed center at 500 mb prior to cyclogenesis along the Atlantic Coast. The upper-level trough was associated with a primary surface low pressure center along the Kansas-Missouri border that was associated with a widespread region of precipitation. The large upper-level vortex drifted slowly eastward in the following 48 h and deepened as the primary surface low weakened and a secondary cyclone formed off the Southeast Coast and then moved to a position just off southeastern New England by 0000 GMT 14 January.

Two remaining trough systems are difficult to place into the previous three categories. The December 1960 case is similar to the category whereby an open wave trough forms a closed center at 500 mb as rapid cyclogenesis occurred along the East Coast. However, this trough was a little more complicated in that it was initially a cut-off low over the southwestern United States on 10-11 December that "opened up" late on 11 December as it propagated eastward (see Fig. 16). This system then merged with a separate open wave trough moving southeastward from central Canada toward the Great Lakes and formed a large vortex across the northeastern United States late on 12 December. The January 1966 upper-level trough is also difficult to place in the simple categorization. In this case, an east-to-west oriented vortex extending from the Great Lakes to off the New England coast split into two separate vortices by 1200 GMT 29 January. One vortex drifted to the east of southeastern Canada, while the other vortex over the Great Lakes rotated cyclonically as a short wave trough to its west propagated southeastward from the Rocky Mountain states toward the central United States. The trough over the Great Lakes appeared to merge with an open wave trough propagating eastward

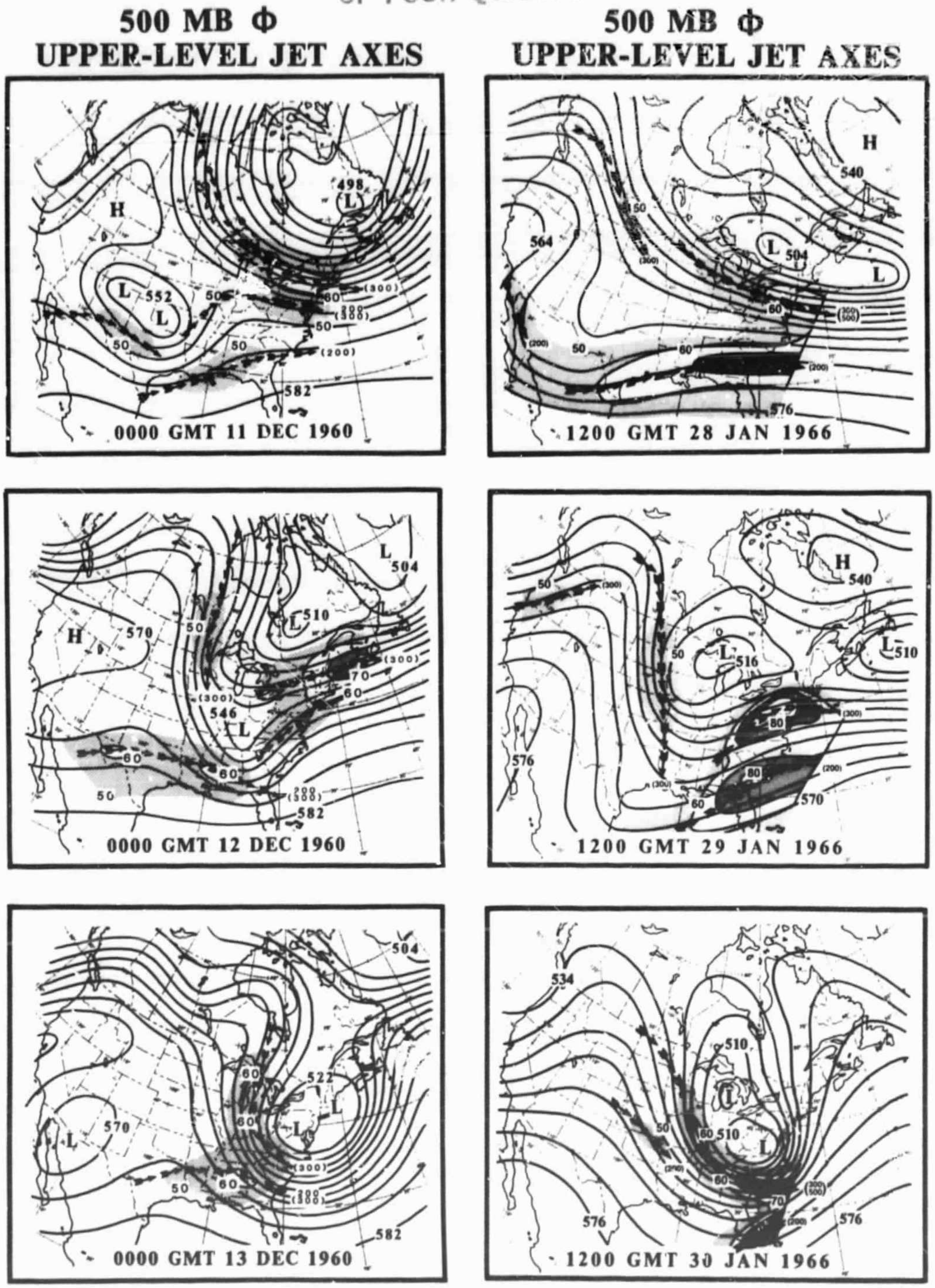

Fig. 16. The 500 mb geopotential height and upper-level wind analyses for two trough systems that are difficult to classify. (Left) 0000 GMT 11, 12 and 13 December 1960, and (right) 1200 GMT 28, 29 and 30 January 1966. See Fig. 13 caption for details.

across the southern United States on 29 January. A vortex center then reformed along the Middle Atlantic coast and deepened as explosive sea-level cyclogenesis commenced on 30 January. Both of these cases were accompanied by the formation of large 500 mb vortices and large increases of amplitude in the trough and downstream ridge during rapid surface cyclogenesis.

a2. Trough/Ridge Amplitude and Wavelength

In order to discuss case-to-case similarities and variations of upper-level features, such as amplitude and wavelength, with respect to a similar stage in the evolution of every storm, it becomes necessary to define a reference point that allows such a comparison, both in space and time. Since rawinsonde observations are taken only once every 12 h, the choice of a reference point that is common to every storm system is difficult to obtain. However, as shown in Fig. 5, there exists a given synoptic reporting time (referred to hereafter as time T) in which all surface low pressure centers are located relatively close together off the Middle Atlantic coast between 34° and 39°N, and 78° and 72°W. All analyses were chosen such that the top-right panel of the 6-paneled figures in Part 2 corresponded to this time (which was 36 h into the 60 h study period). In general, many of the surface low pressure centers at this time and location were deepening, and twelve of eighteen cases were undergoing rapid cyclogenesis (central sea-level pressures were falling at a rate of -3 mb $(3\ h)^{-1}$ or greater). In addition, snow was falling throughout the Middle Atlantic states and the northeastern United States in all cases with moderate to heavy snowfall occurring in several instances.

The amplitudes of the troughs and their flanking ridges were examined to provide a measure of the spatial and temporal evolution of the trough systems during the cyclogenetic period. The amplitudes were determined by taking selected geopotential height contours at 500 mb (they are drawn at 60 m intervals in NMC analyses) and computing the latitudinal displacement (in degrees) between the trough base and ridge crest along a given contour. The maximum displacements between the trough base and upstream and downstream ridge crests were computed for each case to yield a measure of the maximum trough development at 500 mb. The magnitudes and times of occurrence (with respect to time T) of these amplitudes are noted in Table 4.

The amplitudes of the trough and upstream ridge typically reach a maximum between times T and T+12, a period in which the surface lows are deepening rapidly along the East Coast. There is also a considerable case-to-case variation in the maximum amplitudes between the trough and upstream ridge, ranging from 12° (the maximum latitudinal displacement of height contours between trough and ridge axes) for the February 1979 "Presidents' Day" storm to 41° for the highly amplified February 1978 snowstorm, with an average maximum amplitude for the eighteen cases of approximately 23°. The amplitude between the trough and upstream ridge appeared to undergo a significant increase (greater than 5° in latitude) prior to time T in nine of the cases (Table 4). In the category where 500 mb open wave troughs evolved into vortices, five of the eight cases were characterized by an increase of amplitude of the trough and upstream ridge prior to this period. Of the four open wave systems, only one increased in amplitude prior to time T, while two of four closed-contour systems also increased in

amplitude. The increase in amplitude typically occurred as an upper-level ridge across the western United States surged poleward under the influence of strong warm advection along its western flank. As the ridging occurred in the western United States, short wave troughs "dig" southward along the eastern flank of the ridge.

The amplitudes of the trough and downstream ridge typically increase at a later time than increases of the trough and upstream ridge, with a maximum usually reached around time T+24 (Table 4). The maximum amplitudes of the trough and downstream ridge are, in general, smaller than those between the trough and upstream ridge, averaging 15.5°, and are less variable from case to case as well, varying from a minimum of 7° for the "Presidents' Day" storm to 28° in January 1966. During the period of rapid sea-level intensification, an increase in amplitude between the trough and downstream ridge was observed in all but one case (the February 1979 "Presidents' Day" storm). In these instances, the amplitude between the base of the trough and ridge crest increased generally by 5° to 15° latitude. Increases of 10° latitude or more were observed in nine cases and occurred primarily in the category of open wave troughs that evolved into closed vortices (six cases). These large increases of amplitude were not observed for the four trough systems that remained as open waves and were observed for only one case in the closed-contour category. Of the two cases with relatively weak sea-level development (January 1978 and February 1983), neither exhibited a significant increase in amplitude prior to the period when the surface low reached the East Coast. During the period when these systems were producing heavy snow along the East Coast, amplitude

increases occurred, but were less than most of the other cases, averaging near 5^{o}.

To further illustrate the similarities and variations of amplitude (and wavelength as well) that characterize the trough systems at time T, with cyclogenesis occurring off the Middle Atlantic coast, a single 500 mb height contour that extends from the upstream to the downstream ridge was selected for each case (chosen to cross the location of the surface low at time T in Fig. 5) and displayed in Fig. 17. For the type "A" storms, the trough axes are located along a line from the Great Lakes to the southeastern United States (Fig. 17a). The location of the trough axes for the type "B" cases (Fig. 17b) is more spread out across the central and eastern half of the United States, while the small sample of "AB" cases shows the greatest variability (Fig. 17c).

As the amplitudes of the trough systems increased prior to or during cyclogenesis, the half-wavelength between the trough and <u>downstream</u> ridge appeared to decrease in every case (Table 4). In general, the greatest decrease in wavelength occurred during rapid cyclogenesis. However, large decreases in wavelength were also noted prior to rapid cyclogenesis, most notably in March 1960, January 1961, January 1964, February 1979, and April 1982. Mullen (1983) notes that the decrease of the half-wavelength of the thermal fields associated with upper-level trough-ridge systems may be an important clue for non-linear interactions during rapid cyclogenesis. Uccellini et al. (1984) show the importance of decreasing wavelengths between the trough and downstream ridge in enhancing upper-level divergence at the crest of the downstream ridge prior to the February 1979 "Presidents' Day" storm. The increasing upper-level divergence was

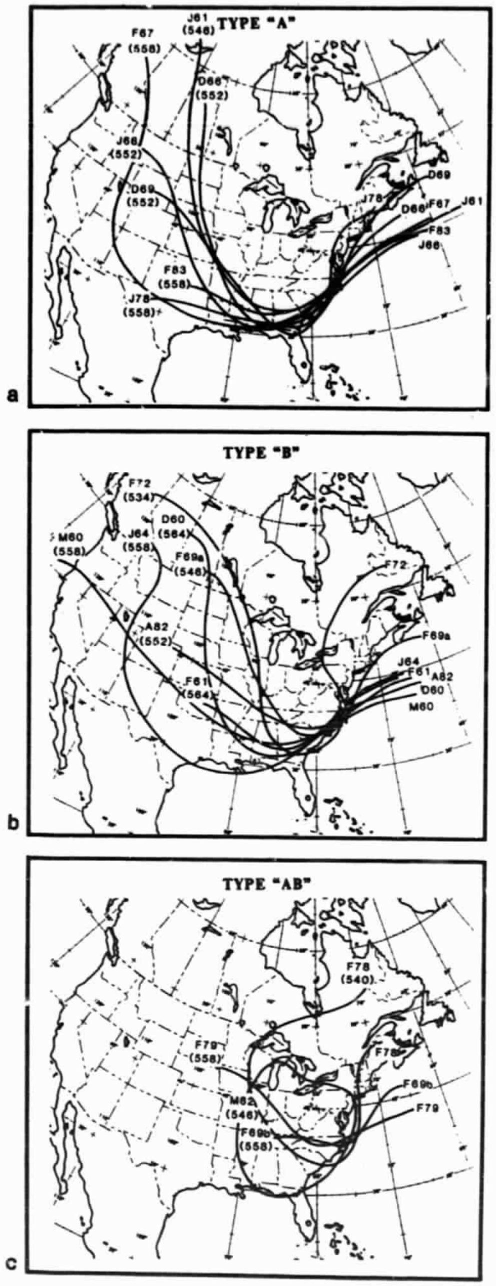

Fig. 17. Illustration of the case-to-case variations of trough/ridge amplitude and wavelength using a selected 500 mb height contour [grouped according to Miller's (1946) classification]. Contour was chosen because it passed near the position of the surface low pressure center at time T (the synoptic reporting time when the surface low center was located near the Middle Atlantic coast). Contours extend from the upstream ridge to the trough axis to the downstream ridge crest. Contour values are in parentheses (564 = 5640 m).

associated with an expanding outbreak of snowfall prior to cyclogenesis and occurred as the coastal front rapidly developed along the Southeast Coast. A similar expansion of precipitation coverage with the onset of coastal frontogensis and cyclogenesis occurred during a period of shortening wavelength in the March 1960 and January 1964 cases as well.

a3. Other Persistent Features

Two persistent features in the 500 mb trough/ridge systems in nearly every case of East Coast snowstorms were the presence of diffluence downwind of the trough axis and a negatively tilted trough axis (Table 4). As pointed out by Bjerknes (1951b) and Palmen and Newton (1969), diffluence aloft is an important signature of increasing upper-level divergence, which significantly influences the surface cyclogenesis (Bjerknes and Holmboe, 1944). The cases in this survey indicate that diffluence aloft is observed primarily during the period of rapid surface development, especially in the exit region of a propagating jet streak as it moves toward the coastline (Section 3b). Diffluence appears to develop downwind of the trough axis, especially in cases where the amplitude of the trough increases significantly. Concurrent with the appearance of diffluence, the trough axis typically assumes the negative tilt from northwest to southeast and is associated with a localized increase of the geopotential gradients at the base of the trough. The negative tilt and increasing diffluence tend to increase the positive vorticity advection and upper-level divergence near the developing surface cyclones. As shown in Fig. 18, the trough axes for these storms, which are spread over a large portion of the central United States 24 h prior to cyclogenesis along the East Coast (Fig. 18a), typically assumed a northwest to southeast negatively tilted configuration

500 MB
TROUGH AXES

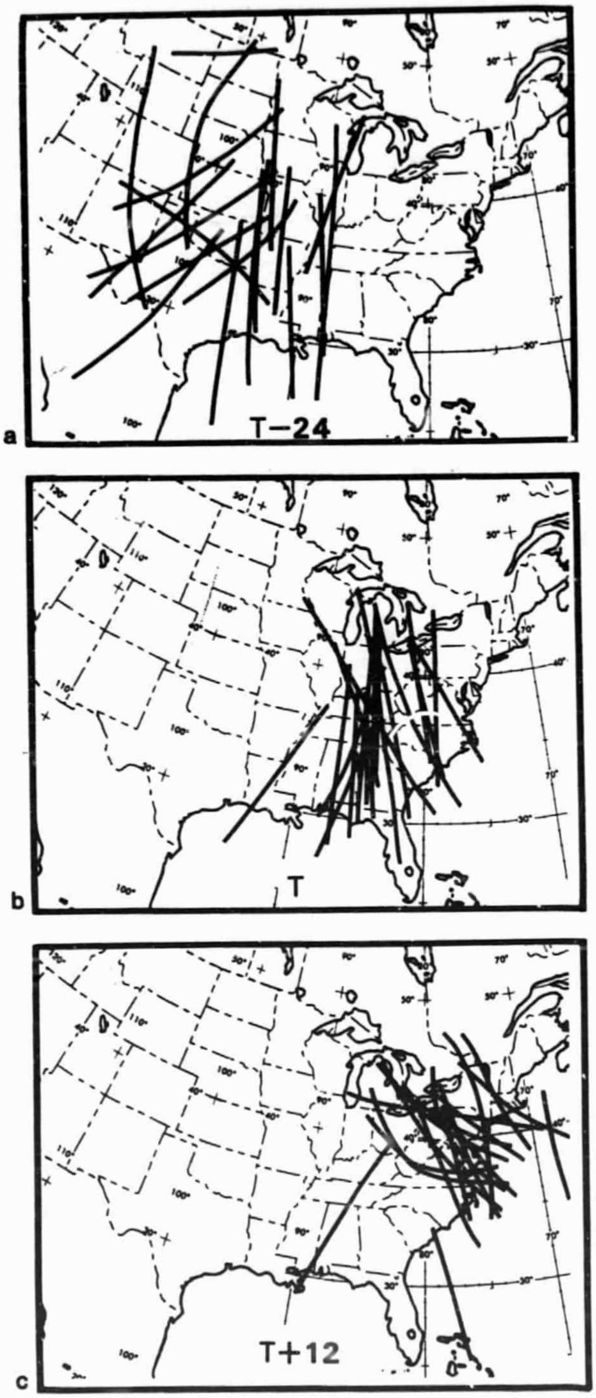

Fig. 18. Axes of the 500 mb trough systems 24 h prior to the development of a surface low along the Middle Atlantic coast (T-24; top), at time T (middle) and 12 h after time T (bottom).

either near time T (Fig. 18b) or more apparently during the 12 h period following time T (Fig. 18c). The period when the upper-level troughs became negatively tilted generally characterized the times when the surface low was deepening rapidly and moving from the Middle Atlantic coast to New England.

Another interesting aspect of the 500 mb flow is the appearance of a well-defined region of confluent geopotential heights over the northeastern United States and eastern Canada which appears to influence the evolution of cold anticyclones during the precyclogenetic period, as noted previously by Reiter (1963). The confluence is located upwind of a trough axis in eastern Canada in nearly every case (see Table 4). During the 24 h period prior to cyclogenesis along the Middle Atlantic coast, distinct height minima that define the centers of trough systems at 500 mb were observed primarily across Quebec, Newfoundland, and Nova Scotia in fifteen of the eighteen cases (Fig. 19a). Confluence was typically located along the axes of the upper-tropospheric jet streak systems that were associated with these troughs and covered much of the upper midwestern and northeastern United States. In general, a strong surface anticyclone and low-level cold air advection across the north-central and northeastern United States were associated with this pattern. By the time cyclogenesis was occurring along the Middle Atlantic coast at time T (Fig. 19b), the height minima had progressed eastward to extreme eastern Canada and the adjacent Atlantic Ocean and confluence was now located along the jet axes which were predominantly oriented across the northeastern United States. Snow was falling generally along the upstream and southern flanks of these jet axes. The presence of the trough in eastern Canada was one of the most persistent

**300 MB JET AXES
ASSOCIATED WITH
EASTERN CANADIAN/NORTH ATLANTIC
TROUGHS**

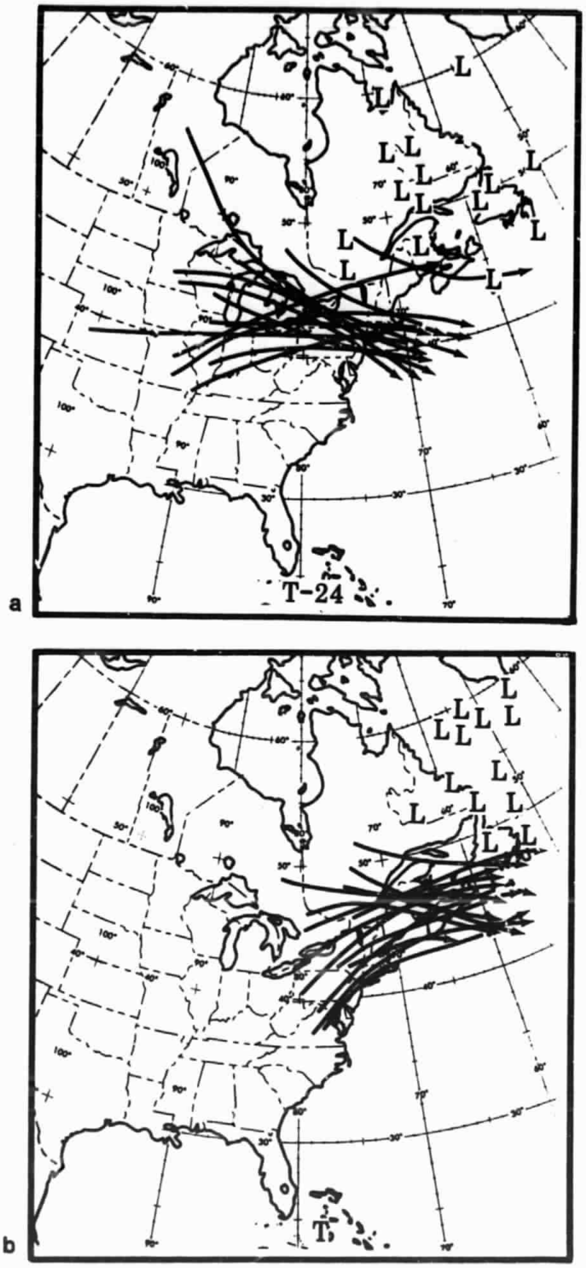

Fig. 19. Summary of height minima (denoted by L) associated with eastern Canadian/North Atlantic 500 mb trough systems and associated 300 mb jet axes 24 h prior to time T (top) and at time T (bottom).

upper-level features of the study. In addition, a ridge at 500 mb was anchored over northeastern Canada or Greenland (e.g., see Figs. 13 and 14) in eight of the cases, impeding the progression of troughs, thus maintaining the confluence over southeastern Canada or New England. In the two cases where snow changed to rain across the Northeastern urban centers (December 1969 and February 1972), a strongly confluent field was absent across southeastern Canada or New England.

Finally, there are cases in which intense East Coast cyclones appear to occur during significant long-term transition states of the atmosphere in which blocking patterns are established (Colucci, 1984). For example, during the three-day period in which the February 1978 storm developed (Fig. 20), the upper-level flow pattern changed from a nearly zonal flow pattern prior to 5 February to one dominated by meridional flow on 6 February, and finally to one where a large, circular and relatively stationary cut-off features developed by 7 February. The intense cut-off upper-level low associated with this intense cyclone meandered throughout eastern Canada during the remainder of February 1978, providing a persistent flow of dry, cold weather to the northeastern United States. Thus, it appears that this cyclone was associated with a new blocking regime that affected the weather across part of the United States for almost a month. As another example of the association between the cyclones and changes in the large-scale circulation, the February 1979 "Presidents' Day" storm signaled the end of a two-week period dominated by persistent, extreme cold and was followed by unseasonably mild conditions (Foster and Leffler, 1979; Bosart, 1981). These two examples are presented to indicate that these storms can be associated with processes that may alter the long

500 MB HEIGHTS

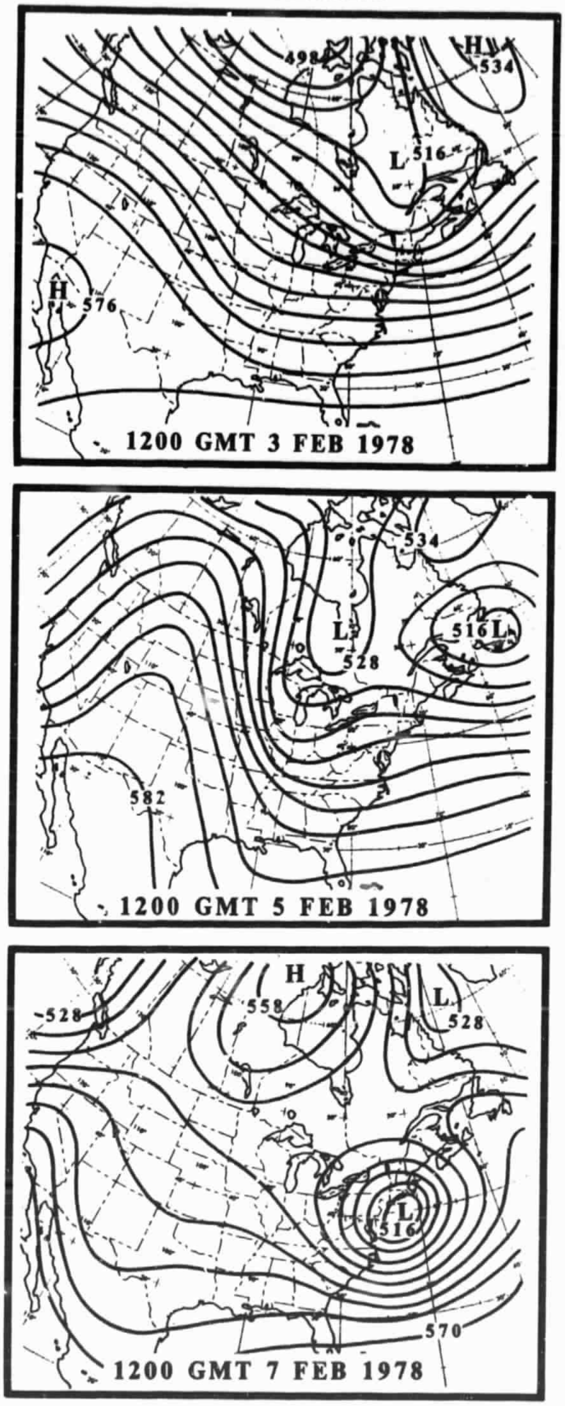

Fig. 20. Illustration of the 500 mb flow pattern (height contours are solid lines, 516 = 5160 m) undergoing a transition from nearly zonal flow (1200 GMT 3 February 1978) to meridional flow (1200 GMT 5 February 1978) to a highly cut-off flow regime (1200 GMT 7 February 1978) during the development of an intense East Coast snowstorm.

wave structure for days or weeks, as is discussed by Palmen and Newton (1969; see Sections 6.4 and 11.5). Further research is needed to determine to what extent these cyclones were responsible for altering the general circulation.

In summary, a common signature of upper-level troughs associated with intense East Coast snowstorms is the development of a closed center at 500 mb. These troughs typically amplify with time and are associated with major sea-level development. Troughs such as these, which comprised approximately half the cases in the sample, were found for both type "A" and "B" systems. In cyclones that did not exhibit a sustained period of rapid intensification, there was correspondingly little change in the trough configuration aloft as open wave trough systems did not develop a closed center at 500 mb. These troughs also did not exhibit the large changes in amplitude that characterized most of the other cases. A final category was the slow-moving vortices that preceded and accompanied rapid East Coast cyclogenesis, which included rather large systems that moved slowly. Three of the four troughs making up this category amplified only slightly during East Coast cyclogenesis, although one case appeared to undergo the largest increase of amplitude prior to and during cyclogenesis of all the cases in the sample.

While marked case by case differences were clearly apparent with respect to the structure and evolution of upper-level trough systems, several features were common to many cases, especially during the rapid surface development phase of the cyclone. These features include an increase of amplitude and shortening half-wavelength between the trough and downstream ridge, diffluence immediately downwind of the trough axis, and

the rotation of the trough axis to a northwest to southeast orientation. Another significant upper-level feature was the presence of a strongly confluent height field across the northeastern United States or southeastern Canada, which was upwind of a trough located in eastern Canada and downwind of the cyclone-producing trough in the eastern United States. The association between confluence and the cold surface anticyclone across the northeastern United States has been established. Its association with upper-level jet streak circulations and influence on snowfall production will be examined in the next section.

b. Middle- and Upper-Tropospheric Jet Streaks

Jet streaks imbedded within upper-level trough/ridge systems enhance and focus upper-level divergence, stratospheric extrusions, vorticity advection, energy exchange and momentum transport, all of which play important roles in cyclogenesis (Bjerknes, 1951a; Reiter, 1963; Danielsen, 1966, Palmen and Newton, 1969; Bleck, 1974; Hovanec and Horn, 1975; Sechrist and Dutton, 1979; Uccellini et al., 1984, 1985).

All cyclones included in this study were associated with one or more jet systems of subtropical and polar origin, as observed from 200, 300, and 500 mb analyses. Polar jet streaks, which were found generally between 300 and 500 mb, attained wind speeds typically between 50 and 70 m s^{-1}. A few trough systems were associated with vast jet systems with maximum wind speeds exceeding 70 to 80 m s^{-1}. Wind speeds associated with subtropical

jet streaks (STJ) approached 80 to 100 m s^{-1} in some instances[2], greatly exceeding the wind speeds of nearby polar jet (PJ) systems. Examples of jet streak systems are shown in Figs. 13 through 16, which illustrate that cyclones may be associated with many jet streak systems at several levels in the atmosphere, with complex interactions among polar, subtropical, and low-level jets (LLJ) (discussed in the next section). For example, in the case of the February 1979 "Presidents' Day" storm, no fewer than four separate upper-tropospheric jet streaks, in addition to an intense LLJ, influenced the development of the storm (Uccellini et al., 1984).

Three persistent characteristics of upper-level wind systems were observed during the study of the middle and upper troposphere during the cyclone events. First, there were several instances in which an upper-level jet intensified near, or slightly upstream of, the base of a trough, with the surface low developing in the diffluent exit region of this jet. The wind speeds associated with the jet at the base of trough were observed to increase by at least 15 m s^{-1} over a 24 h period in seven cases (Table 5), primarily during periods of rapid sea-level development. These wind increases occurred as the geopotential height gradients intensified at and upwind of the base of the trough. Nearly all cases were associated with localized increases of height gradients at 500 mb during cyclogenesis. An example of this is shown for the February 1983 snowstorm (Fig. 14) in which the wind speeds at the base of the trough in the

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[2]As the wind speeds increase in the upper troposphere, the wind measurements become less accurate due to errors related to the low elevation angle (e.g., see Newton and Persson, 1962). Thus, it is difficult to determine if the STJ actually attained those large "measured" wind speeds.

Table 5
Summary of Upper-Level Wind Characteristics

	Increases of Wind Speeds (>15 m s^{-1}) over 24 h or less at the Trough Base	24 h Increases of Wind Speed (>15 m s^{-1}) over 24 h or less in the Downstream Ridge	Polar Jet Streaks Across Northeastern United States or Southeastern Canada during 24 h Period Prior to time T
M 1960	X		X
D 1960		X	X
J 1961	X		X
F 1961	X	X	X
M 1962			
J 1964		X	X
J 1966	X	X	X
D 1966			X
F 1967	X	X	X
F 1969a	X		
F 1969b			X
D 1969			X
F 1972			X
J 1978	X		X
F 1978			
F 1979		X	X
A 1982			X
F 1983	X		X

Table 5. Upper-level wind characteristics, including cases in which increases of wind speed (>15 m s^{-1}) along an upper-level jet streak occurred over a period of 24 h or less at the trough base and in the downstream ridge; and the presence of polar jet streaks across the northeastern United States or southeastern Canada associated with the upper-tropospheric confluent region during the 24 h period prior to time T.

southern United States increased by greater than 20 m s^{-1} over the 24 h period in which the storm was moving up the East Coast. A second characteristic of upper-level winds is the occurrence of an intensifying STJ with maximum wind speeds at 200 or 300 mb. In seven cases, a jet streak amplified near the crest of the ridge immediately downstream of the cyclone-producing trough (Table 5). The amplification of the jet streak near the ridge crest typically occurred prior to rapid cyclogenesis. An example of the STJ amplification (Fig. 16b) shows the development of 80 m s^{-1} jet streaks across the Middle Atlantic and Southeast United States by 1200 GMT 29 January 1966, prior to the movement of an intense sea-level cyclone up the East Coast. The formation or amplification of the jet streaks in this case may occur in a manner similar to that described by Uccellini et al. (1984), in which air parcels at upper levels accelerate near the ridge as they move from the trough axis toward a ridge crest whose half-wavelength is decreasing with time. The amplifying jet systems are usually marked by a swath of upper-level clouds with a distinct, anticyclonically curved edge, as shown for the February 1979 case by Uccellini et al. (1984). The third upper-level wind characteristic was the association of a PJ with the confluent height configuration observed across the northeastern United States and southeastern Canada downwind of an upper-level trough across eastern Canada (Fig. 19). Fifteen of eighteen cases were characterized by a PJ extending across New England prior to or during East Coast cyclogenesis (Table 5). Prior to cyclogenesis along the East Coast (at time T-24, Fig. 19a), the jets were typically directed from the upper Midwestern or Great Lakes states eastward into the Middle Atlantic states and New England. As the cyclone developed along the East

Coast at time T (Fig. 19b), the upper-level troughs across eastern Canada had reached the Atlantic Ocean. The jet axes were now located from the Middle Atlantic states east-northeastward into New England and southeastern Canada, placing much of the northeastern United States in the entrance region of the jet streaks at this time. The appearance of the PJ within the confluent height field is consistent with Namias and Clapp's (1949) relationship between confluence and jet stream formation.

The survey of the eighteen East Coast snowstorms reveals a persistent orientation of upper-level jet streaks associated with the diffluent trough immediately upwind of the surface cyclones and with the confluent flow across the northeastern United States. To illustrate how this orientation of jet streaks can combine to influence the development of heavy snowfall along the East Coast, a more detailed analysis of the jet streaks is shown for the February 1983 storm, which appears to typify other cases as well. The 300 mb analysis at 1200 GMT 11 February (Fig. 21a) shows that the axis of a short wave trough extends from the Ohio Valley to the southeastern United States coast, with an upper-level jet streak centered near the base of the trough over northern Alabama and Georgia. The exit region of this jet streak is located over the Middle Atlantic states and is marked by diffluence[3]. Confluence over the northeastern United States marks the entrance region of a polar jet streak analyzed within the downstream ridge near Nova Scotia.

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[3]The diffluence is shown by the nearly southerly wind report at Washington-Dulles Airport (IAD) and the westerly wind report at Charleston, S.C. (CHS).

1200 GMT 11 FEBRUARY 1983

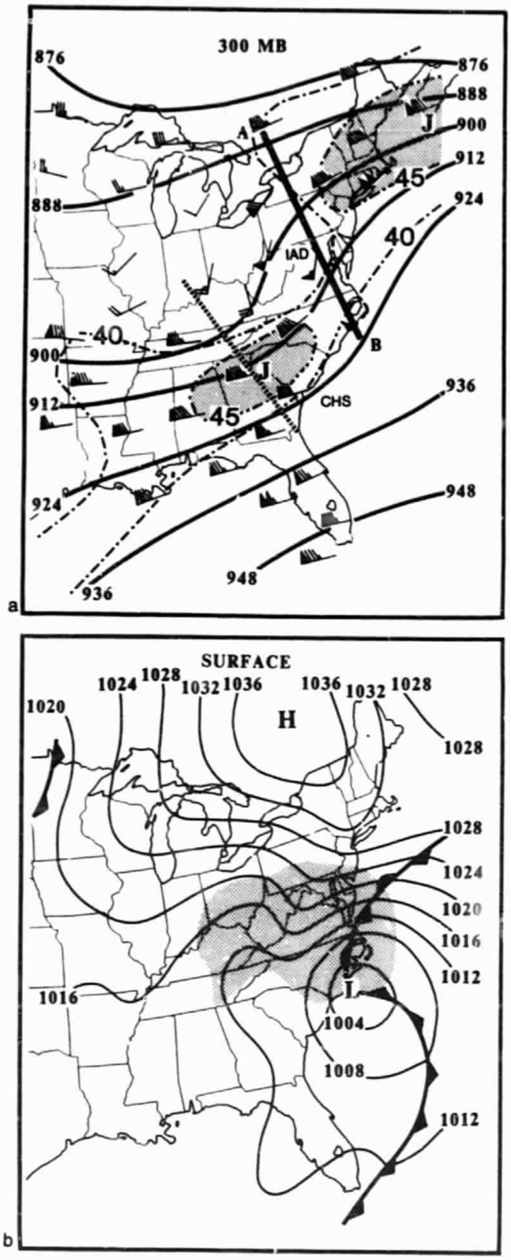


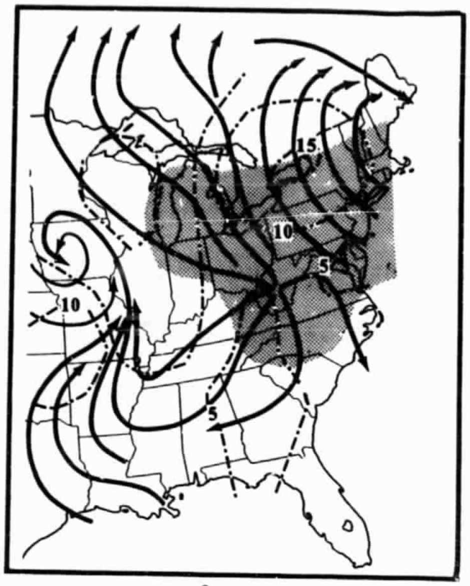
Fig. 21. (a) 300 mb geopotential height analysis (solid, 888 = 8880 m) and isotachs (dot-dashed, 40 = 40 m s^{-1}; each flag denotes 25 m s^{-1}; each full barb denotes 5 m s^{-}; each half-barb denotes 2.5 m s^{-1}; shading represents speeds in excess of 45 m s^{-1}; J denotes locations of jet streaks) at 1200 GMT 11 February 1983. Thick solid line represents axis of cross section shown in Fig. 22 and dotted line represents axis of 300 mb trough. (b) Surface frontal and isobaric (solid, mb) analysis at 1200 GMT 11 February 1983. Shading represents precipitation occurring at the time of the analysis.

At the surface (Fig. 21b), a wedge of high pressure and very cold air over the northeastern United States is located beneath the upper-level confluent region in the entrance region of the upper-level jet streak centered over Nova Scotia. The surface low near the North Carolina coast is located beneath the diffluent exit region of the upper-level jet streak centered over the southeastern United States, immediately downwind of the trough axis. A widespread area of precipitation is located between the two jet streaks, with heavy snow falling across Virginia, Maryland, and West Virginia.

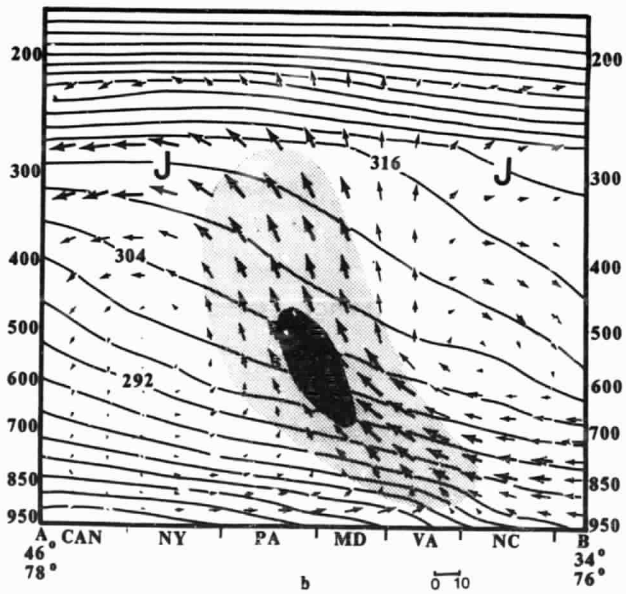
The location of the heavy snowfall in the left front quadrant of the exit region of the polar jet in the southeastern United States and right rear quadrant of the polar jet in the northeastern United States suggests that the rising branches of indirect and direct transverse circulations, respectively, could both be contributing to the ascent needed for the development of such an extensive area of heavy snowfall, as previously suggested by Uccellini (1976). The orientation of the ageostrophic flow in the entrance and exit regions of finite-length, straight jet streaks has been described by many authors, including Bjerknes (1951a), Riehl et al. (1952), Newton (1959), Uccellini (1976), Uccellini and Johnson (1979), Shapiro and Kennedy (1981), and others. In the entrance region of the idealized jet streak, the ageostrophic motion is directed toward the cyclonic side of the jet and represents the upper branch of a direct transverse circulation that converts available potential energy into kinetic energy for parcels accelerating into the jet. The direct circulation is also marked by rising (sinking) motion on the anticyclonic (cyclonic) side of the jet streak. In the exit region of the jet, the

ageostrophic components are directed toward the anticyclonic side of the jet and represent the upper branch of an indirect transverse circulation that converts kinetic energy to available potential energy as parcels decelerate upon exiting the jet. The indirect circulation is typically marked by rising (sinking) motion on the cyclonic (anticyclonic) side of the jet streak. However, Uccellini et al. (1984) discuss how the rising branch of the indirect circulation could actually be displaced toward the anticyclonic side of the jet streak due to curvature effects and the slope of the frontal zone associated with the jet streak.

At 1200 GMT 11 February 1983, an analysis of the ageostrophic wind at 300 mb (Fig. 22a) is consistent with the idealized model of ageostrophic motion associated with straight jet streaks. The cross-contour ageostrophic flow approaching 15 m s^{-1} was directed toward lower heights across the northeastern United States in the confluent entrance region of the jet streak located near Nova Scotia and represents the upper branch of a transverse circulation. Meanwhile, the ageostrophic flow over portions of the Middle and South Atlantic states, especially over Virginia and North Carolina, was directed toward higher heights in the exit region of the jet streak over the southeastern United States and represents the upper branch of an indirect circulation. The east to northeasterly ageostrophic winds in Kentucky, Tennessee, northern Alabama, and Mississippi reflect the subgeostrophic winds found in association with the cyclonic curvature in the trough. The 300 mb divergence was diagnosed across the Middle Atlantic states where the ageostrophic flow regimes diverge from each other, with maximum divergence approaching 6 x 10^{-5} s^{-1} centered over West Virginia. The location of the divergence is also consistent with the positive



a



b

Fig. 22. (a) 300 mb ageostrophic wind streamlines, isotachs (dot-dashed; m s^{-1}) and divergence (shaded) at 1200 GMT 11 February 1983. (b) Vector representation of vertical motions and ageostrophic wind components tangential to the plane of the cross section shown in Fig. 21a, including isentropes (^{o}K) at 4^{o}K increments. The horizontal vector components are scaled at the bottom of the figure (m s^{-1}). Shading represents ascent in excess of -4 μb s^{-1}; heavy shading represents ascent in excess of -8 μb s^{-1}. J's denote axes of upper-level jet streaks.

vorticity advection downstream of the trough axis where the diffluence was maximized.

To demonstrate the effect of the juxtaposition of the transverse circulations in the diffluent exit region of the jet streak at the base of the trough and in the confluent entrance region of the streak located near the downstream ridge axis, vertical motions and ageostrophic wind components were evaluated along the plane of a vertical cross section (Fig. 22b). The cross section was constructed from immediately north of Lake Ontario to the North Carolina coast at 1200 GMT 11 February 1983 (Fig. 21a). The vertical motions are derived by vertically integrating the continuity equation in pressure coordinates using the O'Brien (1970) method. The ageostrophic winds (tangential to the plane of the cross section) are interpolated to the cross section from a 2° latitude by 2° longitude gridded analysis based on the Petersen (1979) objective analysis scheme.

The cross section defines a distinct two-cell circulation pattern oriented normal to the axes of both of the jet streaks. In the entrance region of the PJ across the northeastern United States, the tangential ageostrophic wind components at upper levels are directed toward the cyclonic side of the jet, while the low-level return ageostrophic flow is directed toward the anticyclonic side of the jet. Subsidence is diagnosed in the coldest air immediately above the surface anticyclone in southern Canada. A sloping pattern of ascent, with maximum values of $-8\ \mu b\ s^{-1}$, is diagnosed in the relatively warmer air where isentropes are sloped downward toward higher pressure. The return branch in the lower troposphere contributes to the advection of cold air from New York toward Virginia.

The increasing magnitude of the lower-tropospheric ageostrophic winds from Pennsylvania to Virginia indicates that damming effects east of the Appalachians appear to occur within the lower branch of the direct circulation. This increase is also a significant factor that contributes to accelerating the flow of cold air southward along the coast. This two-dimensional diagnosis of ageostrophic and vertical motions defines the direct transverse circulation in the confluent entrance region of the polar jet across the northeastern United States.

The cross section also displays an indirect circulation in the exit region of the jet streak at the base of the trough over the southeastern United States. In the upper troposphere, the tangential ageostrophic wind components are directed to the anticyclonic side of the jet. The pattern of ascent is now located in the relatively colder air and upper-level subsidence is located in the relatively warmer air with respect to the position of this jet streak. The lower branch of the indirect circulation is characterized by a 15 m s^{-1} southerly ageostrophic wind component directed up the sloping isentropic surfaces generally between 700 and 950 mb, which is equivalent to the diagnosis of strong warm air advection in this region when viewed from the isobaric framework (Sanders and Bosart, 1985a,b; Bosart and Sanders, 1985). The southerly return branch of the indirect circulation appears to influence the transport of moisture from the Atlantic coastal area toward the region of heavy snow in Virginia and Maryland. The characteristics of the lower branch of the indirect circulation and its associated moisture transports are similar to that described for severe weather events by Uccellini and Johnson (1979), and for the "Presidents' Day" storm by Uccellini et al. (1984).

The two circulations combined to produce a sloped pattern of ascent over Maryland and Virginia, in which warm, moist air from the Atlantic and Gulf Coast regions was forced over much colder air moving southward from Canada within the lower branch of the direct circulation. The area of maximum ascent over Maryland and Virginia corresponds to the area where heavy snow was observed at 1200 GMT 11 February 1983. This case illustrates that the presence of jet streaks (one associated with an upper-level diffluent trough in the southeastern United States, and the other with a strongly confluent region in the northeastern United States) can induce a broad, sloped region of ascent that is associated with heavy snowfall. Furthermore, the lower branches of the circulations enhance thermal and moisture transports, creating a more favorable environment for the development of heavy snowfall.

A schematic is shown in Fig. 23 to summarize the configuration of transverse circulations associated with upper-level jet streaks imbedded in a trough and downstream confluent zone that characterize a typical scenario for heavy snowfall along the East Coast: 1) a well-defined trough is located over the Ohio and Tennessee Valleys with a jet streak at the base of the trough entering a diffluent region downwind of the trough axis; 2) another upper-level trough is centered over southeastern Canada with a jet streak imbedded in a confluent zone across New England; 3) indirect and direct transverse circulations are located in the exit and entrance regions of the streaks, respectively; 4) the rising branches of the transverse vertical circulations in the exit and entrance regions of the two jet streaks are associated with a widespread region of clouds and precipitation between the diffluent exit region of the trough and confluent entrance

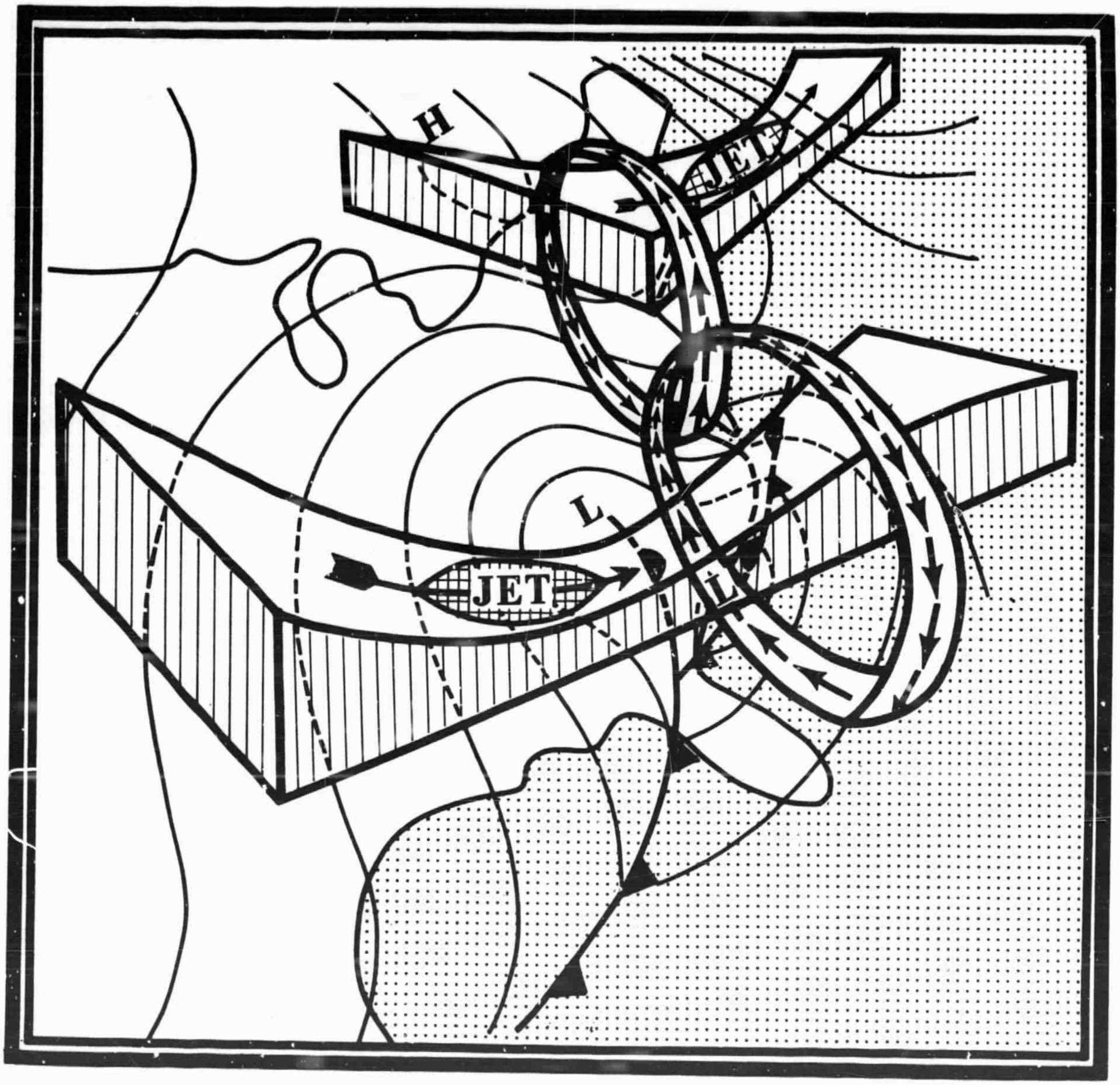


Fig. 23. Schematic of sloping transverse circulations associated with the diffluent exit and confluent entrance regions of jet streaks imbedded, respectively, at the base of troughs moving across the Ohio and Tennessee Valleys and across southeastern Canada that typically occurs during major East Coast snowstorms.

region located further downwind; 5) the advection of cold Canadian air southward in the lower branch of the direct circulation across the northeastern United States provides the thermal support for snowfall; 6) the northward advection of warm, moist air in the lower branch of the indirect circulation across the southeastern United States is forced to ascend over colder air, resulting in precipitation; and 7) the combination of the differential vertical motions in the middle troposphere and the interactions of the lower branches of the direct and indirect circulation may be highly frontogenetic, maintaining and perhaps increasing the thermal gradients during cyclogenesis. It should also be noted that although the transverse circulations are depicted on a two-dimensional vertical plane (Fig. 22b), the circulations are related to three-dimensional variations in the ageostrophic winds and divergence patterns (Fig. 22a) that cannot be defined in terms of simple two-dimensional straight-line jet streak dynamics.

c. Lower-Tropospheric Trough Patterns, Temperature and Wind Fields

Characteristics of lower-tropospheric conditions during the East Coast storm events are examined from 850 mb charts presented in Part 2. Since lower-tropospheric thermal and moisture advections are factors that influence precipitation formation, as well as whether the precipitation falls in liquid or frozen form, the 850 mb surface provides a means to examine these processes. The 850 mb surface is typically examined during snowstorm events since heavy snowfall frequently occurs to the left of the path of the 850 mb low center, and since the $0^{\circ}C$ to $-2^{\circ}C$ 850 mb isotherms often mark the demarcation zone between rain and snow (Bailey, 1960; Browne

and Younkin, 1970; Spiegler and Fisher, 1971). Spiegler and Fisher (1971) claim that the track of the 850 mb low center makes a better guide than the surface low to determine the demarcation zone between rain and snow since its path appears to be more continuous than that of the surface low, in which redevelopment may occur at a distance from the primary surface low center. However, the review of the eighteen cases shows that for the type "B" cases, the motion of the 850 mb low center is not continuous, especially in instances where strong surface redevelopment occurs.

Ten cases (Table 6) exhibited either dual 850 mb low centers or one center with two separate regions of maximum height falls, with the secondary centers forming near the East Coast. All ten cases were associated with surface low pressure centers that are classified as either type "B" or "AB." Dual 850 mb low or height fall centers were observed generally at only one observation time, indicating that the 12-hourly separation of rawinsonde measurements is apparently not able to adequately resolve the rapidly changing evolution of the 850 mb trough (in one type "B" case, no secondary 850 mb low was observed, but it may not have been resolved by the observing network). Dual and discontinuous tracks (Fig. 24) are clearly shown for five type "B" cases (March 1960, January 1964, February 1969a, February 1972, and April 1982) and one indeterminate ("AB") case (March 1962). The figure also shows that the paths of the 850 mb low centers were located slightly poleward of their sea-level centers as they neared the East Coast, which is expected for highly sloped, baroclinic systems (Bjerknes and Holmboe, 1944). The majority of 850 mb low centers moved eastward off the Virginia coast.

Table 6
Summary of 850 mb Characteristics

	Dual 850 mb low centers or height fall centers	850 mb northwesterly flow/ cold advection over the northeastern United States prior to East Coast cyclogenesis (24 to 48 h prior to time T)	Maximum observed wind speeds ($m\ s^{-1}$) and direction associated with 850 mb low center along the East Coast
March 1960	X	X	30(ESE)
December 1960		X	25-30(NE)
January 1961		X	30(SSW)
February 1961	X	X	30(SE)
March 1962	X	weak	35(E)
January 1964	X	X	30(E)
January 1966		X	35(SE)
December 1966		X	30(E)
February 1967		X	30(S)
February 1969a	X	weak	25-30(E)
February 1969b	X		30(SE)
December 1969		weak	30-35(SE)
February 1972	X		30(SE)
January 1978		X	25-30(E)
February 1978	X	X	35(E)
February 1979	X	X	25(SE)
April 1982	X	X	25-30(NE,NW)
February 1983		X	20(NE)

Table 6. The 850 mb characteristics, including the occurrence of dual 850 mb low centers or height fall centers; whether northwesterly flow and cold advection over the northeastern states preceded East Coast cyclogenesis; and maximum observed wind speeds ($m\ s^{-1}$) and direction associated with the 850 mb low center along the East Coast.

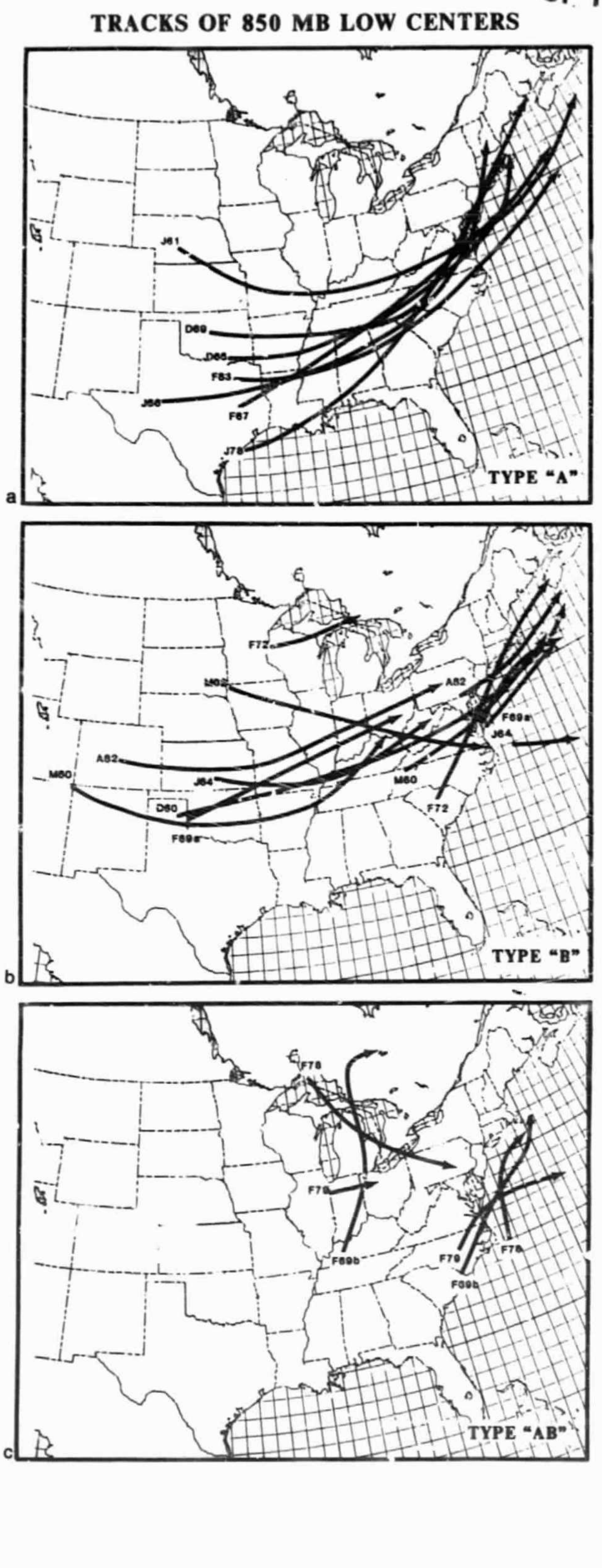


Fig. 24. Tracks of 850 mb low centers over the 60 h study period, grouped according to Miller's (1946) scheme.

A prominent feature at 850 mb during the 24 to 48 h preceding cyclogenesis along the East Coast was strong northwesterly flow and cold advection across the northeastern United States (Table 6) to the rear of an intense 850 mb low center across southeastern Canada. Strong advections and northwesterly flow were observed in at least thirteen of the eighteen cases, in which the $0^{\circ}C$ isotherm was driven south of the Northeast urban centers toward Virginia. As discussed in the previous section, the low-level northwesterly flow was located beneath a region of upper-level confluence in the entrance region of upper-level jet streaks (Table 5), and is likely a manifestation of the lower branch of a direct transverse circulation. In addition, the damming effect discussed in Section 2c also appears to play a role in accelerating the cold air down the coastline along the eastern slopes of the Appalachian Mountains.

With few exceptions, these East Coast snowstorms developed or propagated along a concentrated ribbon of 850 mb temperature gradients, with the surface low located along the warm edge of the gradient. The surface low was also located either at or upwind of the apex of the 850 mb thermal ridge. The gradient of 850 mb temperatures increased during cyclogenesis and formed an "S"-shaped pattern as the cyclones neared the East Coast, when cold and warm advections surrounding the 850 mb low were maximized. The $0^{\circ}C$ isotherm appeared to progress into the Middle Atlantic states in all cases, and was typically located near the 850 mb low center. This isotherm did not move any further northward despite the presence of strong warm advection, implying that cooling associated with strong ascent was offsetting the effects of intense thermal advections.

An important element of cyclogenesis, especially along the East Coast, is the development of a pronounced low-level jet streak (LLJ), frequently noted on the 850 mb analyses. With the onset of rapid cyclogenesis at the surface and at 850 mb, the LLJ typically develops from a southerly to easterly direction (Table 6) and acts to augment lower-tropospheric thermal and moisture advections toward the area of heavy snowfall. The LLJ was observed in all of the cases and associated wind speeds were typically on the order of 25 to 35 m s^{-1} (Table 6). The development of these wind maxima near the 850 mb level during the East Coast storm period likely accounts for the secondary maximum of LLJ's along the North Carolina coast noted in Bonner's (1965) survey of low-level jets in the United States.

While the LLJ typically develops in association with the deepening of the 850 mb vortex, it may also form near the coastal zone in advance of the development of the main 850 mb low center, as occurred for the February 1979 "Presidents' Day" storm (Fig. 25a). The development of the LLJ (with wind speeds approaching 25 m s^{-1} over Georgia and South Carolina at 1200 GMT 18 February) occurred in a region of significant 850 mb height falls displaced a considerable distance from the main 850 mb trough. In this case, the LLJ was found to be a result of a three-dimensional adjustment process where upper-level divergence and sensible and latent heating all contributed to the rapid acceleration along the axis of the lower-tropospheric wind maximum (Uccellini et al., 1983). The LLJ was probably linked to coastal frontogenesis off the Carolina coast and enhanced the moisture transport into the developing snowshield in the southeastern United States during the pre-cyclogenetic period (Uccellini et al., 1984).

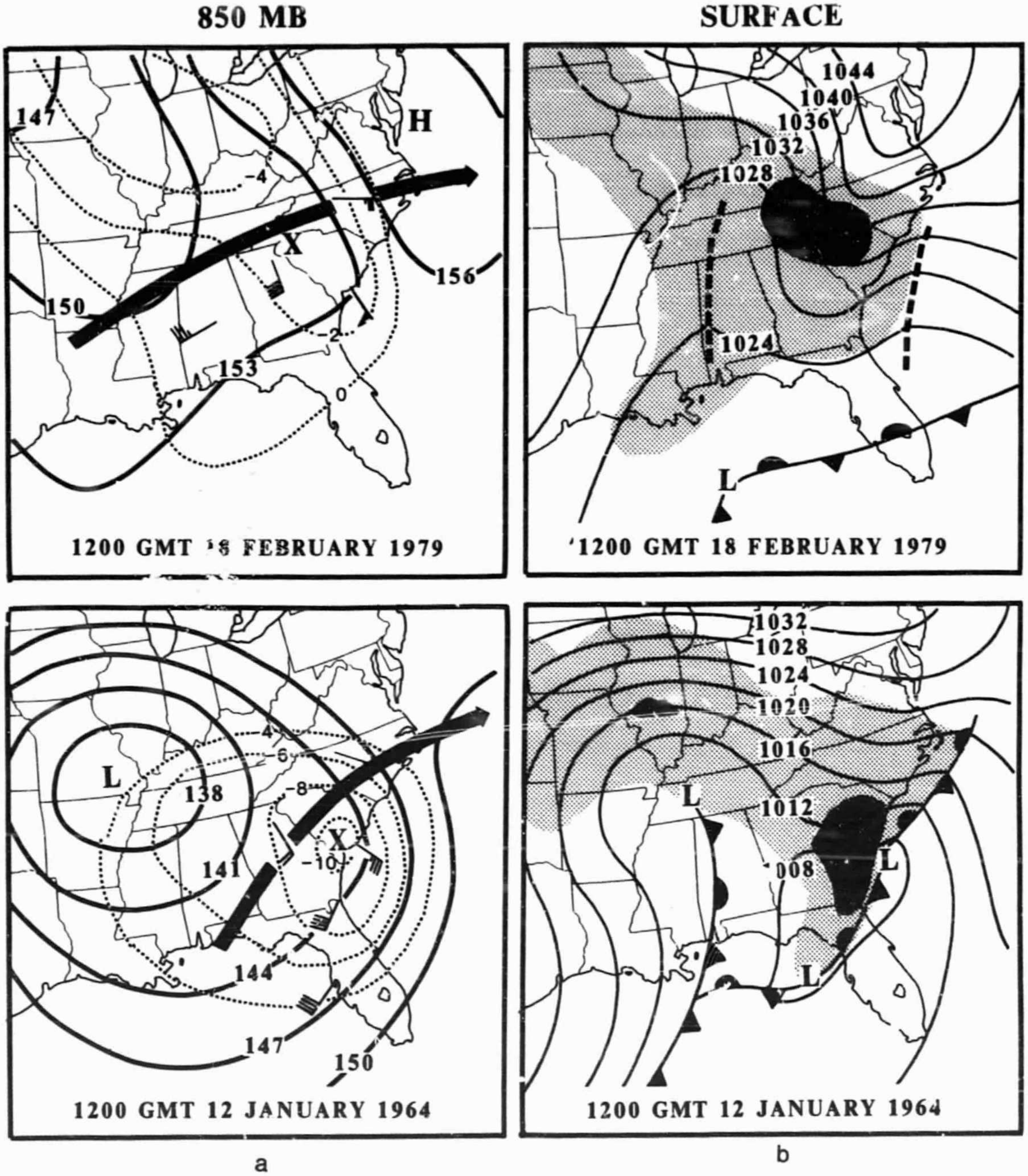


Fig. 25. Two similar cases of 850 mb development at 1200 GMT 18 February 1979 and 1200 GMT 12 January 1964, prior to major East Coast cyclogenesis (see text for details). (a) 850 mb analyses, including wind reports (flags denote 25 m s^{-1}; barbs denote 5 m s^{-1}; half-barbs denote 2.5 m s^{-1}), height (solid, 150 = 1500 m) and height tendency (dotted, -4 = -40 m $(12\ h)^{-1}$) analyses and the axis of the subtropical jet (dark arrow). (b) Surface analyses. See Figure. 3 caption for details.

Several other cases displayed similar characteristics in the 850 mb height and wind fields, especially with regard to the redevelopment of a coastal height fall center displaced from the main 850 mb troughs for the type "B" cases (Table 6). One case in particular was remarkably similar in many aspects to the aforementioned "Presidents' Day" scenario. The January 1964 snowstorm was associated with the development of a noticeably ageostrophic 850 mb LLJ in combination with a coastal height fall center displaced approximately 1000 km from the main 850 mb low (Fig. 25b). As with the "Presidents' Day" storm, the LLJ for this case formed during a period when the upper-level STJ amplified between trough and ridge axes that were decreasing in wavelength, and heavy precipitation and coastal frontogenesis were occurring along the Southeast coast (Fig. 25b). Therefore, the dramatic meteorological changes along the Southeast coast prior to the onset of the Presidents' Day snowstorm were unusually dramatic, but not an isolated case.

4. SUMMARY

An analysis of the surface and upper-level patterns of selected wind, temperature and pressure fields for a sample of eighteen severe East Coast snowstorms indicates that there are large variations in surface and upper-level features prior to and during cyclogenesis. This variability (not only of spatial features, but in their temporal aspects as well) makes the use of quantitative composite analyses very difficult. However, there are persistent features that appeared for many of the cases. A very persistent signal was the presence of rapidly deepening surface lows along the Atlantic Coast. Seventeen of eighteen cases maintained a deepening rate of -1 mb $(h)^{-1}$ or more for at least 12 h. In approximately half of the cases, a secondary surface cyclone formed along the Middle Atlantic coast (type "B"; Miller, 1946) as an initial, or primary, low pressure center remained to the west of the Appalachian Mountain range. The development of the secondary low pressure center appears to be strongly influenced by the mountain range and modulated by influences of land-sea effects including cold air damming and coastal frontogenesis. The secondary cyclonic development was also apparent at 850 mb, with the appearance of dual 850 mb low centers or height fall centers. The primary surface low pressure centers, in general, did not deepen rapidly or did so for short periods prior to or during the period that the secondary surface low formed. The primary low then filled once the secondary low began to deepen rapidly. In a few of the cases in which no secondary low formed (type "A"), the surface low pressure center had a tendency to "jump," or reform, further downwind along the East Coast. A few cases had

characteristics of both "A" and "B." Coastal frontogenesis and cold air damming were also persistent features that were associated with the development of the storms in many of these cases.

A distinct upper-level trough was observed for every case. Therefore, no cyclone evolved under the conditions of straight-line flow postulated by Petterssen and Smebye (1971), where cyclonic systems develop first from the surface and then aloft. Three general categories of 500 mb troughs associated with the surface low pressure systems were discussed. Approximately half of the eighteen cases were associated with "open wave" troughs that evolved into a closed vortex during rapid surface cyclogenesis. The second category described open wave troughs that did not evolve into a closed vortex at 500 mb (four cases) and was associated with cyclones that did not deepen as rapidly as those in the previous category. The final category described troughs with closed centers at 500 mb prior to and during East Coast cyclogenesis (four cases). In all of the trough/ridge systems studied, some form of amplification at 500 mb was involved. For example, the amplitude of the <u>upwind</u> <u>ridge</u> and trough increased prior to East Coast cyclogenesis in half of the cases. However, the amplitude of the trough and <u>downwind</u> <u>ridge</u> increased during cyclogenesis in all but one case. In addition, the half-wavelength between the trough and downwind ridge appeared to decrease in every example, mostly during periods of rapid sea-level development, but also prior to rapid cyclogenesis in a few cases. The development of diffluence downwind of the trough axis, a negative tilt of the trough axis from northwest to southeast, and an increase of the geopotential height gradients at the base of the troughs were also features common to many storms. The rapid

development phase of the cyclogenesis usually commenced beneath the pronounced diffluent region to the east of the upper-level trough axis as the axis became negatively tilted.

Multiple jet structure of polar and subtropical origin also appear to have considerable influence on the development of East Coast storms. As the upper-level troughs developed the pronounced diffluent region to the east of the trough axis, the surface lows appeared to intensify in the exit region of an upper-level jet streak. In general, at least half of the cases were characterized by localized increases in wind speeds prior to and during cyclogenesis. Three general characteristics, two involving amplifying jet streaks, were observed from this study. First, a jet appeared to develop (or increase in magnitude) at the crest of the ridge downstream of the cyclone-producing trough in seven cases during the 12 to 24 h period prior to cyclogenesis along the East Coast. Satellite imagery and surface reports indicate that an expansion of clouds and precipitation area appear to have accompanied this development. Uccellini et al. (1984) found that the large increase of wind speed near the ridge crest downstream of the trough axis prior to the "Presidents' Day" storm was associated with an increase of upper-level divergence and contributed to ascending motion over the region of significant snowfall. A second general characteristic is a jet amplification observed at the beginning of rapid cyclogenesis in seven cases in association with an increase of the geopotential height gradients at the base of the cyclone-producing trough. A third persistent feature that appeared throughout the study was the presence of a jet streak entrance region within a pattern of 500 mb confluent geopotential heights across the United States and southeastern Canada prior to and during East

Coast cyclogenesis. The confluence was typically located upwind of a major trough across eastern Canada. The confluent entrance region of an upper-level jet streak across the northeastern United States and the diffluent exit region of a separate upper-level jet streak at the base of the cyclone-producing trough nearing the East Coast were studied during the recent February 1983 snowstorm and indicated that the production of heavy snowfall was related to the juxtaposition of transverse circulations associated with these jet systems. This configuration of jet streaks and its associated circulations appears to be a common feature in many storms (see also Uccellini, 1976), and has several effects. The rising branches of the two circulations combine to produce a broad, sloping region of ascent. The lower branch of the direct circulation located in the confluent entrance region of an upper-level jet streak across the northeastern United States is also responsible for enhancing lower-tropospheric cold advection (in addition to the effects of cold air damming), thus providing a mechanism for maintaining air along the coastal regions of the East Coast that is cold enough for snowfall. The lower branch of the indirect circulation in the diffluent exit region of an upper-level jet streak at the base of the cyclone-producing trough is responsible for enhancing warm advection and moisture transports toward the region of heavy snowfall. The juxtaposition of jet streak circulations, therefore, appears to be a crucial element for producing heavy snowstorms along the East Coast.

These findings suggest that the combination of jet/trough systems can provide the upper-level divergence/baroclinic instability needed to initiate the development of a cyclone, but that topographical features

related to the East Coast, in particular, land-sea effects, the Appalachian Mountains, and boundary layer processes over the ocean act to focus or modulate the rapid development phase of the surface cyclone. Field experiments being planned (such as GALE and STORM EAST), which are designed to supplement operational observations over land and especially over the ocean immediately along the East Coast, should attempt to resolve how these physical mechanisms interact to produce the heavy snowstorms capable of paralyzing the heavily populated regions along the East Coast.

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15. Supplementary Notes

16. Abstract Surface and upper-level characteristics of selected meteorological fields are summarized for eighteen major East Coast snowstorms dating from 1960 to the present. Two major "types" (Miller, 1946) of sea-level development are described and applied to the cases at hand, with a few storm systems showing characteristics of both types. Aspects such as rapid sea-level deepening, coastal frontogenesis, cold air damming, low-level jet formation, the development of an "S"-shaped isotherm pattern at 850 mb, diffluence downwind of a negatively tilted upper-level trough axis, upper-level confluence across the northeastern United States, and an increase of geopotential heights at the base of the upper-level trough characterized the pre-cyclogenetic and cyclogenetic periods of many of the storm systems. However, large case-to-case variability was also observed, especially with regard to the spatial dimensions of the surface and upper-level systems, as well as variations in trough/ridge amplification and the evolution of upper-level jet streak systems. The influence of transverse circulations associated with a confluent jet streak entrance region in the northeastern United States and the diffluent exit region of a jet streak/trough system approaching the East Coast on the production of snowfall is also discussed.

17. Key Words (Selected by Author(s)) snow, snowstorm, cyclone, extratropical cyclone, forecasting, cyclogenesis, jet streaks, coastal fronts, damming, blizzards	18. Distribution Statement

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